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FOR many years it has been our custom to publish special issues or supplements devoted to some topic of timely interest. In keeping with this tradition, this issue consists of a group of papers on reefs, both ancient and modern. We are indebted to Dr. Heinz A. Lowenstam not only for his contribution but also for his assistance in arranging and editing this number of the *Journal*.

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The next issue of the *Journal* will contain a group of papers on feldspar. Among these are:

"The System $NaAlSi_3O_8$-$KAlSi_3O_8$-H_2O." By N. L. BOWEN and O. F. TUTTLE

"The Alkali-Feldspar Join in the System $NaAlSiO_4$-$KAlSiO_4$-SiO_2." By J. F. SCHAIRER

"Gallium and Germanium Substitutions in Synthetic Feldspars." By JULIAN R. GOLDSMITH

"The Optics of Triclinic Adularia." By URSULA CHAISSON

"The Lattice and Twinning of Microcline and Other Potash Feldspars." By FRITZ LAVES

"High-Temperature Albite and Contiguous Feldspars." By O. F. TUTTLE and N. L. BOWEN

"An X-Ray Investigation of the High-Low Albite Relations." By FRITZ LAVES and URSULA CHAISSON

"On the Relation between Anorthite Content and γ-Index of Natural Plagioclase." By FELIX CHAYES

"Note on 'Pseudo-monoclinic' Plagioclase." By CHRISTOFFER OFTEDAHL

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——— and DUNSANY, A. J. (1929*a*) Early Ordovician faunas of South Dakota: U.S. Geol. Survey Bull. 444, pp. 1–76.
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CAPITAN BARRIER REEF, TEXAS AND NEW MEXICO[1]

JOHN EMERY ADAMS AND HUGH N. FRENZEL
Standard Oil Company of Texas

ABSTRACT

The Capitan is the best known of the larger Permian barrier reefs of the Southwest. This paper describes the sedimentation and depositional environment of this well-known reef.

INTRODUCTION

The Guadalupe Mountain front, rising sharply from the broad plains to the southeast, parallels U.S. Highways 62 and 180 from Carlsbad Caverns in New Mexico to El Capitan Peak in Texas. This bold barrier scarp with its deep transecting canyons, leading down from the dark, tree-capped mountains behind, is the type-section of the Capitan Reef, the youngest and best known of the Permian barrier reefs in the Southwest. Throughout the Permian a deep mediterranean embayment linked the broad, shallow, epicontinental seas of the Permian Basin with the open ocean to the south. In the Capitan stage a continuous barrier reef grew up around the head of this deep channel. This asymmetrical ridge separated the deep seaward waters from the shallow back-reef lagoons. The part of this reef loop that lies north of the Rio Grande is some 450 miles in length. The 45 miles of exposed reef front between Carlsbad and El Capitan Peak is thus only a small segment of the entire Capitan barrier. Additional limited segments are exposed in the Chinati, the Glass, and the Apache mountains, but for some 350 miles of its course the reef is buried by later sediments.

The Capitan was recognized as a fossil barrier reef in the late twenties. Early work by Crandall (1929), Lloyd (1929), and Blanchard and Davis (1929) blocked out the main problems of reef architecture. This work showed that the Capitan was a barrier-reef classic. Study of growth stages, ecology, organic content, reef geography, structural relationships, origin, and sedimentary petrology still yield profitable data. This work is being carried on by geologists with more time to study the outcrops and with a greatly extended knowledge of the buried areas supplied by hundreds of wells drilled for oil. The most complete bibliography of the southwestern Permian reef areas is that supplied by King (1948) in his recent report on the southern Guadalupe Mountains.

PRE-CAPITAN SEDIMENTATION

In the Mississippian period broad areas of the Southwest were covered by epicontinental seas characterized by rela-

[1] Manuscript received November 3, 1949.

tively uniform sedimentation. The early Pennsylvanian was a time of major deformation. Broad structural arches developed first. Slightly later the crests of these arches were pushed up to form narrow mountain chains. Middle and late Pennsylvanian seas overlapped the eroded roots of these mountains. Many of the ranges, however, were rejuvenated by late Pennsylvanian uplifts and continued to act as positive units throughout most of the Permian. The mountains followed an irregular east-west to northwest-southeast pattern that split the Pennsylvanian seas into semi-isolated subrhomboidal basins. Except over local uplifts and isolated peaks these basins were the sites of almost continuous deposition from early Pennsylvanian on into the Permian. In the Pennsylvanian, limestones were deposited in the shallow seas, while shales and sands accumulated near shore and in the deeper basins.

The Permian epicontinental seas reached their greatest extent early in that period, at which time they covered wide areas of the western and southwestern states and of northern Mexico (Hills, 1942). Most of the Permian Basin was only mildly negative, but in northern Mexico, southwestern Texas, and eastern New Mexico it became an active autogeosyncline. The geosyncline extended southwestward as a deep channel to connect with the open ocean. In this southern area subsidence was so rapid that the only shallow-water deposits were the organic reefs and the narrow zones of biostromal limestones deposited along the edges of the emergent mountain chains and on the shallower seamounts. The actively subsiding areas were separated from the mildly negative ones by structural hinge lines. Shoreward of these hinge lines extensive shallow-water marine limestones, that extend back to the outer edge of the near-shore clastic zone, were deposited. Most of these limestones were biostromal.

The deeper waters on the basinward side of the hinge line were divided into several basins by the old, partly submerged Pennsylvanian ridges. Surface waters circulated freely through shallow connecting channels, but deep circulation was limited. As a result, the closed bottoms of these basins were deficient in oxygen. Strong concentrations of CO_2, probably supplied by decaying organisms and kept in solution in the cooler bottom waters, dissolved most of the finer carbonates that were precipitated or washed into these depressions and freed them for redeposition on the reefs or on the broad limestone flats. The normal pontic deposit of these deeper basins is black siliceous silt, although black shales, black limestones, and gray to dark sandstones are also present. Because there was an excess of sulfur, pyrite is relatively common. Most of the fossils are silicified; chert is abundant in the uncommon limestones. Some of the silts are full of radiolaria; others contain abundant siliceous sponge spicules. Lenses of white limestone or dolomite and thin beds of gray sandstone were deposited during periods of freer circulation. Cores from some of the shaly silts show presolidification flowage structures.

The supply of clastics reaching the more active southern pontic basins was never great enough to fill them up level with the surrounding banks. In part, this failure to fill the basin was due to the rapid rate of subsidence, because greater thicknesses of sediments are present in the basins than are found in the surrounding positive areas. The main cause, however, was insufficient clastic material. The Permian Basin was not surrounded by major mountain uplifts com-

parable with those bordering most geosynclines. Neither was it a receiving basin for major silt-laden streams. The mountain ranges within the basin were narrow and were soon cut down and buried. The broad, shallow, limestone seas, surrounding the deeps, served as settling basins to trap much of the land-derived clastics that entered the basin. As a result, differential topographic relief between the tops of the limestone banks and the floors of the pontic basins was maintained or even increased. Maximum relief in the early basins was probably less than 500 feet, and the marginal slopes into these deeps were fairly flat.

The hinge lines at the edges of the limestone banks were favorable places for the growth of incipient reefs. The water was clear and shallow, and sunlight, necessary for plant growth, could penetrate to the bottom. The break in slope forced the cooler, food-packed bottom waters to rise to the surface, and the clastic-free bottoms furnished good points of attachment for fixed organisms.

The first reefs were isolated mounds which soon reached sea level, where upward growth was stopped. Algae were the main reef-building organisms. Waters behind the reef were shallow and warm and had lost much of their CO_2 and other plant foods. A much more abundant food supply was present in the cool waters flowing through the channels between the mounds. To take advantage of this extra food, the reef organisms expanded laterally and eventually formed almost continuous barriers. This resulted in a perched profile of equilibrium. Stratigraphic evidence indicates that the topographic relief of these early marginal reefs did not exceed 300 feet. The flat slopes on the basinward side of the reefs were covered by a broad apron of bedded limestones, in part deposited in place by bottom-living organisms and in part washed off the reefs and back-reef flats as limestone muds.

Because these barrier reefs grew around the head of a mediterranean embayment, they faced seaward in broad concave arcs.

An examination of the pre-Capitan Reef strata suggests the following history. Subsidence was almost continuous. The southern geosynclinal area went down faster than did the marginal flats, but this area of geosynclinal subsidence was progressively restricted as the period advanced. Although the over-all movement in the marginal areas was downward, the sinking was not uniform. In times of normal subsidence, reef growth was concentrated along established zones. Growth was upward and slightly forward. When sinking exceeded upbuilding, the reef was drowned and a new reef zone was established in shallower waters nearer shore. When sinking was interrupted by short periods of uplift, the old reefs were raised above sea level and a new line of reefs was established along the euphotic zone on the old fore-reef apron. Shifts in both directions are recorded, but, because the area of maximum subsidence was decreasing, advances are more common. Each advance resulted in a shelf of lagoonal limestone resting on a pontic foundation. Greater subsidence in the basin tilted the overlapping shelves basinward.

Around the heads of some of the more rapidly subsiding segments of the pre-Capitan deep basins the foreward step-outs were narrowed down, and the reefs grew almost vertically through great thicknesses of sediments. These piled-up reef masses, although not resting on a solid foundation, acted as intermediate positive elements. The Howard-Glasscock bank in Texas and the Maljamar

bank in New Mexico are typical examples.

The early Permian reefs which had started growing on the crests of the old basinal Pennsylvanian banks and on isolated seamounts, such as the deep Big Lake dome, had short histories. The heights on which they grew were located in areas of most rapid Permian subsidence. The bottom went down so fast that the reef organisms were unable to grow upward fast enough to keep in the zone of sunlight. As there were no higher points to retreat to, the reefs were drowned and buried by the black silts.

The reefs along the ridges in the southern geosynclinal basin were more successful. They started as fringing reefs or atolls along the mildly negative mountain crests. The adjacent basin waters were deep from the beginning, so that the reefs grew more steeply upward than did those in the flatter marginal areas. Before the central peaks were submerged, the reefs had grown laterally to form continuous barriers on both sides of the mountain chains. Later these barriers formed the margins of flat-topped platforms, of which the Central Basin Platform and the Diablo Platform are the best-known examples (P. B. King, 1942). Because these reefs grew upward rather than outward, they built a relatively incompressible limestone foundation. Wells and exposures show a few discontinuous sections of pontic rock wedging into the platform margins and a few minor tongues of limestone extending for short distances out into the basins (R. E. King, 1942). In general, however, the platforms with their bordering reefs were products of a stable environment unlike the fluctuating marginal shelves.

The marginal shelves grew basinward until their outer edges coalesced with the shoreward ends of the interbasin platforms. The platforms, as mentioned above, were built of solid limestone resting on a firm, but mildly negative, base. The shelf foundation was a thick section of pontic clastics accumulated in the adjacent, more actively subsiding basins. Differential compaction and more rapid sagging combined to produce slumping and down-folding along the jointure lines. Movement continued throughout the period, and the dip-slopes that mark the margins of the old supported platforms are conspicuous features of all later Permian sediments. This is especially well illustrated along the northeastern edge of the Central Basin Platform in Gaines and Andrews counties, Texas.

The barrier reefs along the outer margins of the platforms and shelves grew up into narrow ridges, with their crests at or near sea level. The area behind these ridges formed a lagoon, comparable in origin to, but greater in extent than, the lagoons within an atoll or between a fringing reef and the shore line. Many of the pre-Capitan lagoons were several hundred miles wide. The floors of the lagoons were shallow and relatively flat. Waters in the lagoons were continuous with those beyond the reef ridge. Along the reef ridge there was an almost constant exchange from one side to the other. The breaking waves and flooding tides moved in over the ridge, and then the piled-up waters moved out again through surge channels and open holes in the reef rock. On the shallow flats, far behind the reef, waves and tides moved the waters up and down and backward and forward for moderate distances, but there was no general interchange of lagoon and basin waters because of this movement.

The Permian is commonly referred to as a period of general aridity. Certainly, the enclosed intracontinental Permian

Basin drainage area seems to have been deficient in rainfall. Water supplied by streams, plus precipitation on the lagoonal and basinal areas, seldom equaled evaporation. As a result, the surface of the water within the basin tended to fall below sea level. The basin and the sea were connected by an open channel. A steady surface current flowed from the open ocean into the basin to maintain both at the same level. This was largely a one-way journey, because the reef barriers and the shallowness of the lagoons hindered any deep outflow. Continuation of this excessive loss and replacing inflow led to the deposition of evaporites.

Precipitation of evaporites began in Lower Permian Wolfcamp time at the inner end of the lagoonal sea in Kansas and eastern Oklahoma (Hills, 1942). Thereafter, evaporites were among the commoner sediments. By middle Leonard time the evaporite seas extended to the outer edges of the lagoons.

Evaporite deposition followed a number of regular laws, which will be discussed briefly. The broad, shallow, flat-bottomed back-reef lagoons acted as huge evaporating pans. The surfaces of these pans were subject to an equal amount of evaporation for every equal unit of area. To maintain the surface of the whole lagoon at sea level, the waters at the inner end of the landlocked lagoon had to be renewed just as regularly as did those at the seaward entrance of the basin. This necessitated a continuous inward flow of water from the reef to the inner margins of the lagoonal pans. As it moved, the water was continuously evaporated, with loss of volume and concentration of the dissolved salts. In time the less soluble salts began to crystallize and settle out of solution. Because precipitation of any one salt took place through a fairly wide range of concentrations and because the waters from which the salts precipitated were moving, the resulting deposit covered a wide area on the lagoonal floor. After precipitation of the least soluble salt, the waters moved on to an area where the precipitation of the next least soluble salt began, and so on. As a result, the precipitated salts were deposited in a lateral rather than a vertical sequence, with all the different salts being precipitated from the same body of water at the same time. There was a fairly wide overlapping of precipitation concentrations at which the successive geologically important evaporite salts were lost from solution, so that along the border lines between zones halite may be mixed with anhydrite, anhydrite with dolomite, and dolomite with limestone. Intergranular mixtures resulted from the overlapping of precipitation concentrations (Adams, 1932).

Changes in sea level caused by storms and vertical shifts along the hinge lines produced interlaminated mixtures. A great influx of storm waters would crowd the evaporite sea to the back edge of the lagoons and temporarily freshen the water in the outer reaches. Negative shifts of the crest of the reef barrier allowed large quantities of the concentrated heavy salt brines to escape down the open seaways. The reversal of these processes tended to shift the evaporite seas outward. Mixing, both intergranular and interbedded, of lagoonal sediments is common throughout the Permian.

Even though the Permian Basin did not receive enough stream water to flush the brines off the lagoonal pans, the streams did supply a large, but not overwhelming, amount of clastics. In the early Permian the protruding mountain crests in the center of the basin supplied thick lenses of coarse conglomerates to the adjacent seas. Later, when these

mountains were buried, clastic sediments were derived from the landward margins of the lagoons. In the early stages, shales, many of them red, interfinger with the offshore limestones and evaporites (Page and Adams, 1940). The highly saline waters along the lagoon margins flocculated most of the clays that came in, and the shale tongues rarely extend far from the shore. A few clastic fingers extend far out across the lagoons. Some may even have reached the back edge of the reef zone. These few widespread clastic wedges enable one to correlate the thick, sparingly fossiliferous back-reef beds.

During this early and middle Permian shale stage silt and sand and some shale were supplied to the central geosynclinal basins. These coarser clastics were not derived from the same sources as were the marginal shales. Perhaps they came from Pennsylvanian and older formations exposed along the seaway connecting the Permian Basin with the open ocean and were deposited in the basins by inflowing currents.

CAPITAN STRATIGRAPHY

The sequence of deep pontic basins–marginal reefs–back-reef lagoons–evaporation pans–near-shore clastics that was initiated early in the Permian continued to evolve throughout the period and culminated in the Capitan Reef and its associated deposits. The Capitan, with its well-balanced physical and organic adaptations, is a natural outgrowth of this long period of sedimentary evolution, modified somewhat by an increased rate of subsidence in the constricted marine areas. At the beginning of Capitan deposition the structural pattern of the southern Permian Basin was well established. The main features are shown on the accompanying map (fig. 1). Additional basins, platforms, and channels probably existed to the south and west, but their outlines are not known. All the positive areas and marginal shelves seem to have been bordered by well-established middle Guadalupian reefs.

There are three stages of Capitan Reef growth. The first of these is a short, youthful stage of establishment and rapid spreading of reefs to enclose the shallow areas. The second is a mature stage dominated by upward growth. The last and shortest of all is a time of decadence and death.

The base of the Capitan is correlated with the Cherry Canyon and Bell Canyon contact of the Delaware Basin, the San Andres and Grayburg contact on the shelves and platforms to the north and east, and the Vidrio and Altuda contact of the Glass Mountains, as indicated in table 1.

The Capitan Reef and the equivalent pontic beds of the Delaware Basin have not been subdivided. The back-reef beds have been split into a number of units that are named, widely recognized, and used. Throughout this paper the back-reef names are extended to express time equivalency in the reef and fore-reef sections.

YOUTH

The first stages of Capitan Reef growth are recorded in the basal Grayburg formation and its equivalents (Skinner, 1946). The sediments above and below the base of the Grayburg show that the Capitan was introduced by a pronounced down-warping of the basinal areas and a corresponding upwarp of the landward margins of the basin. Clastics of the pre-Capitan San Andres formation on the shelves and platforms are largely colloidal shale. Fine sand replaces the shale in the basal Grayburg. This change may indicate a difference in source, but more probably it records a slight steepening of

FIG. 1.—Late Permian structural provinces

the gradient. Bentonite, which is common in the Grayburg, seems to have come from a volcanic area in Mexico.

Most of the older reefs were drowned by the pre-Capitan submergence. Reef organisms quickly re-established themselves in the optimum euphotic zone around the head of the deep seaway. Along most of its course this new reef rests on a bedded limestone deposited in the shallows behind the old reef line. This early Grayburg reef line was very irregular. It followed the winding contours of the pre-Capitan shelves, swinging far back into the re-entrants and side channels and lapping close around the steep faces of the shallow, submerged banks. The approximate position of this initial Capitan Reef zone is shown on the accompanying map. In the Big Bend area of Texas the position of the early reef is unknown. It is deeply buried or completely eroded away, and in the buried areas there are few wells that penetrate the Permian. On the northern and eastern edges of the geosyncline, where the outcrops are well exposed and wells are numerous, the outline can be determined fairly closely.

Maximum depth of the deeper early Capitan basins was probably about 600 feet. These deeps were limited to the Delaware Basin and to connecting basins and channels to the south and west. In the Midland Basin the waters were shallower, and, except in the sand-bottomed areas along the southern and western margins, the basin floor was covered by limestones. The warping that initiated the Grayburg produced some low arches on the shallow basin floor. Isolated reefs are found around or over the higher of these submerged hills. One of these mound reefs in southern Reagan County furnishes the Grayburg reservoir for the Big Lake oil field. Another localized the

TABLE 1

UPPER PERMIAN CORRELATION CHART

	Fore-Reef			Reef Zone		Back-Reef	
	Glass Mountains		Guadalupe Mountains	Glass Mountains	Guadalupe Mountains	Glass Mountains	Guadalupe Mountains
Ochoa	Not present			Not present		Tessey	Dewey Lake Rustler Salado Castile
Upper Guadalupe	Altuda	Delaware Mountain	Bell Canyon	Capitan	Capitan	Tessey Gilliam	Tansill Yates Seven Rivers Queen Grayburg
Lower Guadalupe	Word (western facies)		Cherry Canyon	Vidrio (including Lower Massive Member)	Goatseep	Vidrio	San Andres

World pool in Crockett County. The mounds were built of oölitic dolomite protected by algal ridges. Flank rocks are less dolomitic and contain a rich and varied marine fauna. Most of the Grayburg interreef deposits of the Midland Basin are light-colored dolomites and limestones interbedded with sands. Around the head of the Sheffield Channel these carbonates give way to pontic sands. Waters of the Delaware Basin were too deep for the initiation and growth of basinal reef mounds.

Water circulated freely through the deep seaways in front of the Capitan Reef. A constant current flowed inward across the reef to maintain the waters of the evaporating pans at or near sea level. The reef organisms thus had an adequate supply of food and minerals. Even corals, which are one of the most sensitive groups of marine animals, are fairly common in the reef and fore-reef areas. This would indicate that the waters that reached the reef line at the head of the long seaway were sufficiently like those of the open ocean for corals to thrive.

Reef growth in the early Capitan was rapid and subsidence of the basin slow. As a result, the top of the reef quickly reached sea level and could grow no higher. Competition for space to expand was keen, and, as the richest supply of food was in the seaward waters, growth was most rapid on the front face of the reef. Waves broke off many of the most exposed growths; but, because the food situation remained the same, they were quickly replaced. These broken fragments helped build up a talus slope. Mixed with the wave-broken rubble was fine lime mud washed out of the back-reef areas and the skeletal remains of organisms normally growing on the fore-reef slope or swimming about in the marginal waters. In the shallower re-entrants along the reef front very little talus fill was necessary to bring the bottom up to a depth where the reef could grow across it. In some narrow channels lateral growth near the mouth of the channel closed the gap before the organisms around the head of the channel had had a chance to advance that far. As a result, advances of the reef line in the shallower areas were so fast that the reef may be said to have jumped forward. This process tended to straighten the reef because the reef-bordering deep waters, where a large amount of talus was necessary to build up the bottom, grew forward much more slowly. The advances of the reef line are difficult to map because the age of the bottom beds, both in outcrops and especially in the subsurface, is generally uncertain.

In early Capitan time the Sheffield Channel was an open seaway. By the close of the Grayburg, however, reef growth at its western end had sealed off this 75-mile channel and had thus cut off marine circulation in the Midland Basin. A wide area at the northern end of the Delaware Basin had been overgrown by reefs, and a similar rapid advance of reefs occurred in the western Glass Mountains. By these and other smaller jumps the reef line had been shortened to about two-thirds its original length. This filling-in or cutting-off of the shallow channels and the great shortening and smoothing-out of the reef front mark the close of the first, or youth, state of Capitan Reef development. This point seems to have been reached at approximately the same time along the entire reef front.

MATURITY

In the middle Capitan (Queen, Seven Rivers, Yates, and early Tansill) stage, reef development continued to follow the pattern already outlined, but with sev-

eral outstanding modifications. Most of the early Capitan Reef growth took place on or near the edges of the mildly negative platforms or banks. In the mature stages the reef began encroaching on actively negative basins. In spite of marked subsidence, organic buildup was capable, at all times, of maintaining the crest of the reef at or near sea level. Most of the time subsidence was so slow that the reef grew forward on the foundation supplied by its own talus. In those areas where the supply of food and minerals was at a maximum the advance was especially rapid. One of these areas of greatest step-out was at the northern end of the Delaware Basin, where the reef is now some 30 miles in front of its line of origin. Elsewhere the advance probably averaged less than 5 miles. Because of the greater subsidence under the more advanced areas, such as the northern Delaware Basin, the reef is thicker there than it is in the areas on either side.

Spasms of regional subsidence punctuated the slow, regular sagging of the basin. During and immediately following each such period of down-dropping the reef organisms expended their energy in growing upward to regain their place in the sun. In the intervening quieter periods, after rapid upward growth was checked at sea level, the steady advance was resumed. A series of upbuilding steps and intervening advances is recorded in the walls of McKittrick Canyon, which cuts completely across the reef and back-reef area on the east slope of the Guadalupe Mountains (King, 1948). Much of this part of the section has been eroded from the Glass Mountains, but here also a series of steps and rises is indicated.

Perhaps the subsidence was caused by compaction of the underlying basinal clastics as the heavy reef mass grew upward and forward across them. The bottom-set beds, however, dip uniformly basinward, with no breaks under or in front of the reef. The subsidence that affected the reef is, therefore, the result of substratal movements and not of differential compaction or of local adjustments to load. This differential pressure was fairly high. Finally, the Capitan Reef crest, at the northern end of the Delaware Basin, stood 1,800 feet higher than the basin floor in front. Assuming an average specific gravity of 2.6 for the reef mass and 1.027 for the sea water, the differential was about 1,220 pounds per square inch, or 2,530,000,000 tons per square mile, greater under the reef than under the water-filled basin in front.

The top of the reef was essentially at sea level at all times. The basin floor in front, however, sank steadily farther and farther below sea level because deposition within the basin failed to keep pace with subsidence. The Queen, Seven Rivers, and Yates formations of the lagoonal areas contain bedded sands, many of which can be followed into the immediate back-reef areas. The failure of these sands to build up the basin bottoms faster than they did suggests that the marginal reef ridge was a fairly effective barrier in preventing the passage of lagoonal clastics into the basin.

The Yates is the best known of the sandy formations. Yates time was marked by a widespread sand encroachment that blanketed the lagoonal areas of the Permian Basin. Near the reef the sands are interbedded and cemented with dolomite; farther back they are associated with anhydrite and salt. Sands extend right up to the back edge of the reef wall, that is, to within about a fifth of a mile of the reef face. The thickness of the sandy section in the immediate back-reef area is about 250 feet. In the

lagoons the sands are thinner and less uniform. Yates sands, like those in the preceding Seven Rivers and Queen formations, are normally fine- to very fine-grained, with scattered, large, frosted quartz grains. At the top of the Yates, sand deposition ceased in the back-reef and lagoonal area.

The formation of extensive, thick sand bodies in a section generally free from major sandstones suggests a structural, erosional, or sedimentary change. At the reef edge of the lagoons the evaporites crept closer to the back edge of the reef than in the preceding and following nonsandy stages. This weakens the suggestion that the flood of sand was caused by increased rainfall. Possibly the sands moved across the back-reef flats as dunes. Just in front of the reef many of the sands of the Bell Canyon formation contain scattered, coarse, frosted sand grains like those in the back-reef sands. Apparently they were washed over or through the reef wall. The fact that the reef was not choked by sand is proof of the continuity and vigor of its growth.

The effectiveness of the reef as a barrier to the passage of inflowing waters varied considerably from place to place, but it is probable that at high tide waters passed freely across the crest. Where the reef was backed up against a relatively positive structural element like the Central Basin Platform, secondary asymmetrical barriers, with steep basinward-facing slopes, were sometimes built along the edge of the old platform or along the outer edge of older buried reef lines. There is a tendency for old topographic forms of structural origin to perpetuate themselves through great thicknesses of younger carbonate sediments. Sometimes this is the result of recurrent slight positive movements. In other cases it may be caused by more limestone being deposited over the slightly higher areas than in the adjacent lows. These secondary ridges show up on detailed structural maps of the back-reef areas. They usually served to slow up the movement of inflowing waters. As a result, the limestone-depositing seas rarely extended for more than a mile or two beyond the lagoonward edge of one of these back-reef barriers.

Where the reef grew across an actively negative segment, like the north end of the Delaware Basin and the west end of the Sheffield Channel, with no backup from old secondary ridges, water circulated freely across the barrier. Each down-sagging of the basement opened a series of cross-reef channels for the inflow of fresh waters and perhaps also for the escape of heavy, saturated brines from the evaporite pans. This transfer slowed down as soon as the reef rim again reached sea level. In the meantime, the free influx of unsaturated surface waters furnished the food and carbonates necessary for new reef growths and for a local expansion of the limestone-depositing sea, which, opposite the low areas, sometimes extended for 30 or 40 miles behind the Capitan barrier.

In these highly negative areas, where the reef and back-reef areas were subsiding in a series of almost continuous steps, the waters behind the temporarily ineffective barriers were at times deep enough and fresh enough for the growth of reeflike mounds. These mound reefs are confined to areas just behind the main barrier, where the waters were fresh enough to support numerous colonies of encrusting algae, which, like their relatives of the barrier wall, stood up against the breaking waves. Like other new reef growths, these mound reefs started in the zones of optimum depth. These optimum zones occurred on

the crests and flanks of some of the higher submerged hills in the back-barrier areas. In time the mound reefs grew upward to sea level. Inflowing currents sweeping around or between the mounds kept the bottoms cleared of the fine limestone muds which were piled up in the lee of the reefs. When the barrier reef again reached near enough to sea level to check the rapid inflow, many of the mound-reef algae were killed by the increasing salinity of the waters. Usually, however, before normal back-reef deposition had completely buried the mounds, a new subsidence started them growing again.

Several of these reef mounds are exposed in the hills west of Carlsbad. The outcrops indicate that most of them were limited to the late stages of Capitan Reef growth. Others have been encountered in wells on the Carlsbad Shelf to the east, and some of these are of middle Capitan age.

The mound reefs of the back-barrier negative area appear to have grown actively in the periods between sand floods. As a rule, the sand beds of the Yates are thinner and the iron content is reduced over the mounds, whereas they thicken and remain oxidized in the intermound basins and channels. In the vicinity of the mound reefs and the back edge of the barrier, the dolomite beds between the sand layers are full of pisolites, algae, and other fossils. There is no evidence in the fore-reef beds or at the toe of the reef that great amounts of sand were poured across the barrier crest during the periods of rapid sagging.

DECADENCE

In the late (upper Tansill) stage of Capitan history, sedimentation departed from the previously established pattern. For a time the reef continued to advance and to grow upward as the bottom subsided. Then Capitan Reef growth came to a sudden stop, at least in the Delaware Basin portion of the area, and there was a marked change in sedimentation. To understand the death and burial of the Capitan Reef, it is necessary to know whether the reef was cut off in the prime of maturity, by outside agencies, or whether it passed through a period of old age, natural decline, and death.

The youngest back-reef beds equivalent to the Capitan are the upper Tansill dolomites. The Tansill beds overlie the Yates and average about 100 feet in thickness over wide areas of the back-reef zone (DeFord and Riggs, 1941). In early Tansill time the limestone sea was much wider than in the preceding Yates. This suggests a generally free movement of inflowing waters along the entire reef trend, but the pisolite and fossil zones of the Tansill are much narrower than in the older beds. Most of the rock is made up of finely granular dolomite mud. Anhydrite crystals and inclusions are common in these lithographic beds almost up to the back edge of the reef wall. Both algae and animal fossils are extremely abundant on the reef and for some distance down the fore-reef slope. The absence of fossils in the uppermost back-reef Tansill and the corresponding restriction of the carbonate sea suggest that the lagoons were again creeping up on the reef. This encroachment, however, may have resulted from a limiting of free circulation into the Delaware Basin rather than from local causes. Few other evidences of reef decadence have been noted.

The most logical cause for the killing of the Capitan Reef was the formation of a barrier, presumably of reef origin, across the geosynclinal channel to the south. Pinch-off barriers of this type had

formed several times in the Permian Basin. An excellent example is the growth of the Capitan Reef itself across the west end of the Sheffield Channel to cut off the Midland Basin. In each case the old reef was killed and a new, shortened reef segment was formed. The fact that the first post-Capitan deposits in the Delaware Basin are evaporites supports the idea of a new pinch-off reef barrier.

In most previous pinch-offs, however, the size of the back-reef lagoons was increased by the amount of the area cut off. In the case of the Capitan, the old lagoonal areas were temporarily abandoned, and a new lagoon, restricted to that part of the deep basin which had been cut off, was created. This suggests a slight regional uplift. Possibly uplift, warping, and reef growth went together. The new barrier cannot be definitely located from present data. On fossil evidence the upper-reef limestones of the Las Delicias area of southwestern Coahuila are thought to be younger than any Capitan known in Texas or New Mexico (R. E. King *et al.*, 1944). Thick back-reef-type carbonate in the Tessey formation of the Glass Mountains, which is, in part at least, post-Capitan in age, and the southward "limeing-up" of other post-Capitan formations suggest that the new pinch-off reef barrier grew across the channel connecting the Delaware and Marfa basins. However, no post-Capitan Permian sediments have been recognized in the Marfa Basin, and the barrier may lie farther to the south.

POST-CAPITAN SEDIMENTATION

At the close of Capitan deposition the Delaware Basin was uplifted and tilted slightly eastward (Adams, 1944). The uplifted and tilted depression was filled with back-reef evaporites. Because of the tilt the section along the western margin was somewhat thinner than along the east. When the depression was almost full, there was a renewal of subsidence in the center of the old Permian Basin. The lagoonal sea expanded eastward out of the restricted Delaware depression to reoccupy extensive areas to the north and east. Possibly the western rim was not submerged. The first deposits of this expanded lagoon were anhydrites, followed by a thick section of salt, interbedded, in places, with potash salts and other late-stage evaporites. The salt was followed by a thin layer of anhydrite and limestone, which was in turn buried by late Permian red beds. These impervious, late Permian evaporites reflect the structure of the surfaces across which they were deposited and effectively sealed and preserved the upper surfaces of the buried Capitan.

Post-Permian movements along the buried Capitan ridge at the east edge of the Delaware Basin fractured the overlying late Permian cover. These fractures opened up channels for circulating unsaturated waters to attack the soluble salts. A slump syncline developed. In most cases this slump valley is filled with overthickened Triassic red beds. The slumping reverses the structure of the shallower late Permian rocks but does not affect the Capitan.

Upper Triassic terrestrial clastics apparently covered most of the southern Permian Basin. The Marfa Basin, however, and some adjacent areas were the site of a Triassic-Jurassic highland. Here Jura-Trias beds are missing, and the Cretaceous rests on deeply eroded Permian or pre-Permian rocks. The Cretaceous sea seems to have covered the Capitan on the western rim of the Delaware Basin. Farther east the Cretaceous rests on the uppermost Permian present or on the thin section of Triassic. In the Tertiary

the Marfa Basin again became negative and was buried by thick lava flows extruded from vents to the north and east.

FACIES

A barrier reef is an organic ridge that splits the area in which it grows into two parts, one in front and the other behind the reef. Reef growth starts as a response to sedimentary and structural variations. Once started, the barrier accentuates these initial differences and at the same time introduces an environment of its own. The dominant portion of a barrier reef is the living section between the base of that part of the fore-reef slope where attached organisms control sedimentation and the line on the back-reef flat where fixed organisms cease to control the bottom topography. This organic band may be called the "reef proper," but a sharper dividing line between the fore-reef and back-reef provinces is the crest of the reef wall. The reef wall is therefore a logical place to begin the description of the sedimentary facies associated with a barrier reef.

Reef wall.—In cross section a barrier reef is strikingly similar to an idealized delta, with the reef wall marking the break between top-set and fore-set beds. In life, this reef wall was a self-maintaining ridge that rose high enough above the profile of equilibrium to protect the loosely consolidated materials of the lagoonal platform. In the Capitan the wall was built of organisms that were sufficiently rigid and securely enough attached to each other and to the bottom to withstand the pounding of storm-driven waters and the erosive action of normal waves and currents. The Capitan Reef wall is built almost entirely of the interlocking remains of encrusting algae and calcareous sponges (Johnson, 1942) (pl. 1). These attached lime-secreting forms locked themselves closely together during growth. When exposed portions of the reef were broken or battered by storms, the surviving individuals spread laterally and quickly healed the breaks. Most of the other attached forms present in the reef-wall rock acted as space-fillers rather than as wall-builders. Many of the algae and other fossils of the reef wall have been partially destroyed by dolomitization.

In general, the wall rock is very massive, but in a few places it is penetrated by thin tongues of fore- or back-reef rock. Some apparent bedding in the massive rock is the result of spalling or fracturing. Because the beds in front and behind are almost equally massive, the reef wall is one of the least conspicuous features of the Capitan Reef complex. It is probable that at the time of deposition the top of the wall was higher than the beds on either side, but the absence of bedding makes it difficult to prove this.

During periods of stability, the Capitan Reef edged forward into the basin. The entire front face of the reef, down as far as fixed organisms could grow, was built forward. This forward movement did not take place uniformly along the whole reef front. Some segments moved forward faster than others did. As a result, the reef wall bulges and recedes in an irregular pattern. In shallow re-

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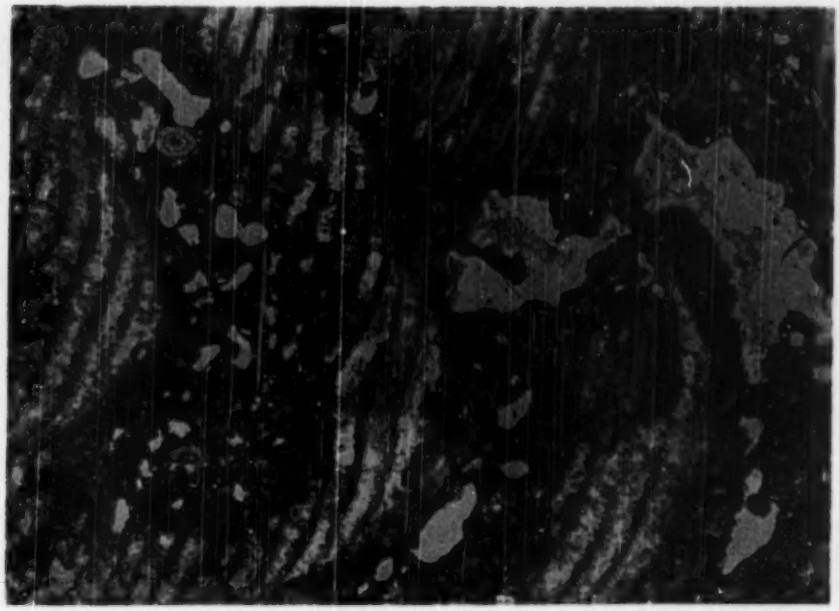
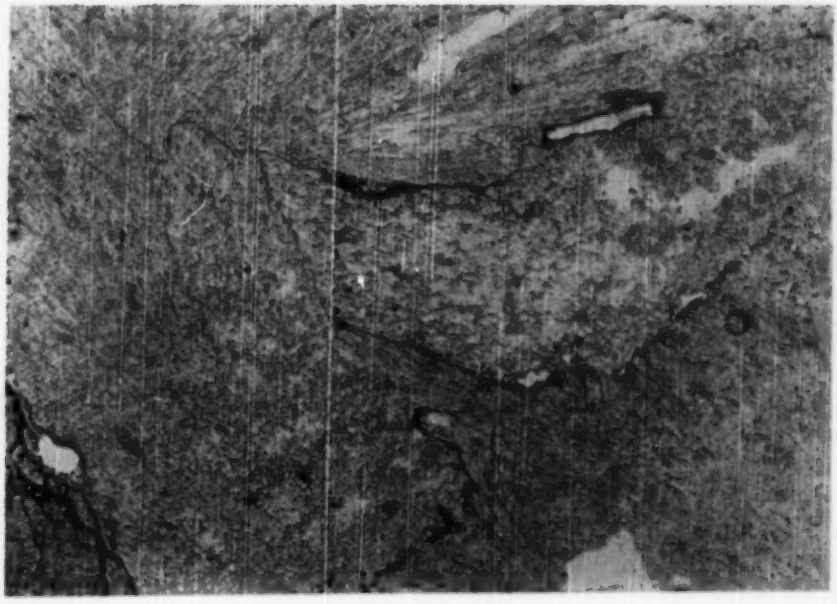
PLATE 1

A, Thin section of upper Capitan reef-wall limestone showing algal growths separated by clastic carbonate debris. From mouth of Dark Canyon, Eddy County, New Mexico. 10×.

B, Thin section of upper Capitan algal reef-wall limestone with filamentous algae and white fluorite inclusions. From reef slope west of White City, Eddy County, New Mexico. 10×.

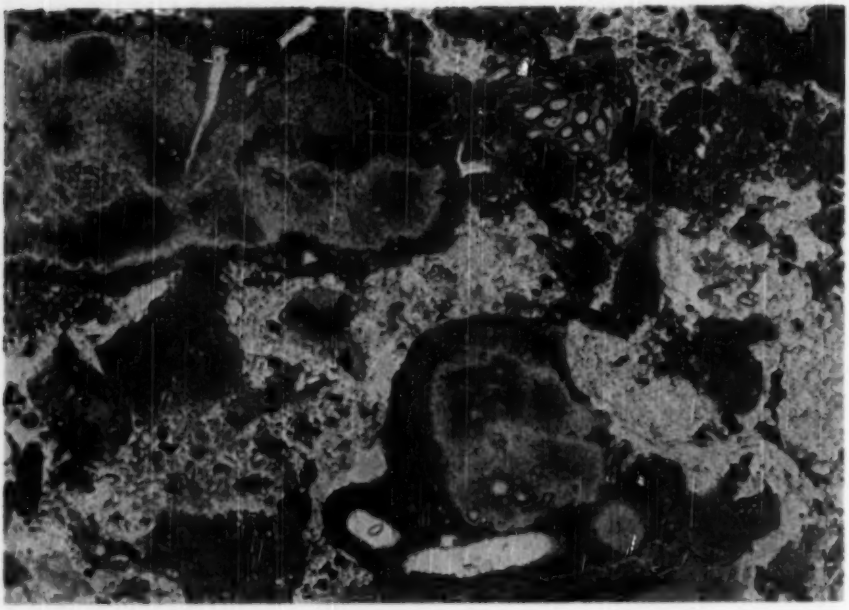


A

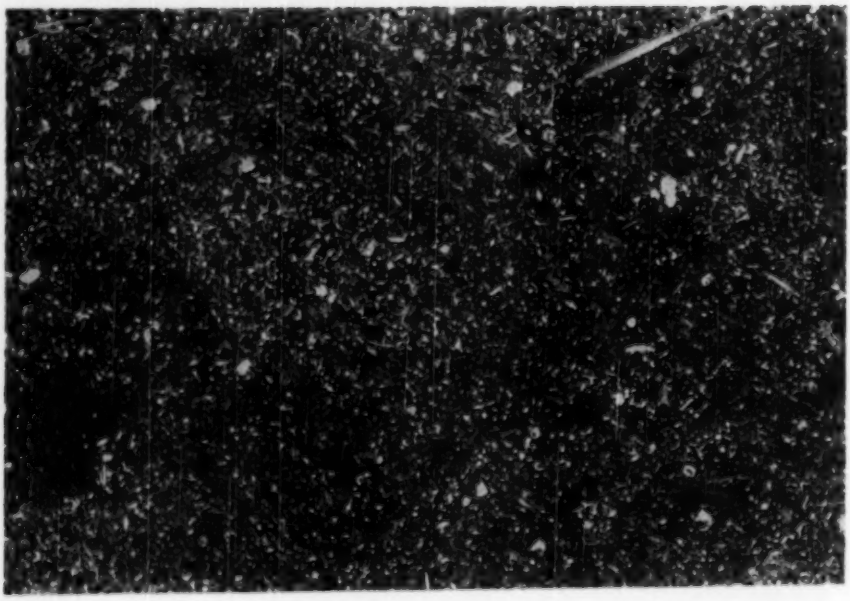


B

Upper Capitan reef wall limestone



A



B

Capitan forereef and basinal limestones

entrants, behind the advanced buttresses, thin beds of reef wall and fore-reef detritus may be interbedded with normal back-reef deposits. When the bottom was subsiding slowly and the reef was building forward rapidly, the reef-wall section was much thicker than the equivalent back-reef section. For example, the average thickness of the Tansill, the highest back-reef formation, is about 100 feet over much of its area of distribution. The thickness of the equivalent reef wall is over 300 feet. Most of this divergence takes place in a short distance behind the wall.

Abrupt sinking checked forward growth of the reef wall and initiated rapid upbuilding. The sections that reached sea level first grew laterally and tended to close the channels, although retreating waves sweeping out through these gaps probably kept some of them open. Some were unfilled at the close of Capitan time. These gaps may have been 100 feet deep or more. The section is too poorly exposed to permit recognition of surge channels, grooves, tunnels, or blowholes. Such features may be present along the buried portions of the reef crest. Almost all the minor drainage lines, along the reef scarp, follow shallow depositional synclines in the face of the reef wall. The gullies referred to here are usually 100 yards or more apart.

Back-reef.—The back-reef Capitan section is divided into four lithologic zones characterized by carbonates, anhydrite, salt, and near-shore clastics. There were wide intergradations between these zones at various times and places.

The carbonate zone extends from the reef where organisms were abundant to a line a few miles back where organisms were absent. Because cross-reef circulation was freer in some places than in others the back-reef zones do not exactly parallel the reef front. Just behind the reef the rocks are mostly poorly bedded dolomites and limestones that interfinger with the reef itself. Many beds consist of pisolites; others consist of filamentous, encrusting, and massive colonial algae. The denser beds are mostly solidified carbonate muds. Much of the carbonate mud originally deposited in this near-reef area was later washed out by retreating waves and currents and redeposited on the fore-reef or even spread across the lagoon floor. Pisolites range from little spheres the size of a pea up to balls 2 or 3 inches in diameter. Many of them are second-generation structures with broken fragments of earlier pisolites for nuclei. Some of the larger balls have multiple nuclei, and there is a complete gradation between the incrustations on these pisolites and the massed algae crusts of the reef wall. The pisolites near the reef are commonly limestone, those farther back are almost pure dolomite, the cement between them is normally coarse-grained calcite. In addition to the algae, Bryozoa, fusulinids and other Foraminifera, large gastropods, heavy-shelled brachiopods, pelecypods, straight cephalopods, holothurians, and other fossils are present near the reef. The average width of the highly fossiliferous zone is less than a half-mile, but it may be wider where the reef wall was broken

PLATE 2

A, Thin section of highly silicified clastic limestone with jumbled fossils and carbonate debris. From the lower fore-reef slope of the Capitan on the south wall of McKittrick Canyon, Culberson County, Texas. 10×.

B, Thin section of dense, dark, basinal-type clastic limestone. From the Lamar limestone ¼ mile east of the mouth of McKittrick Canyon, Culberson County, Texas. 10×.

by gaps. The pisolite zone, which extends considerably farther back, contains encrusting forms responsible for the back-barrier mound reefs. Pisolites are also very common on the secondary barrier ridges that border the more positive back-reef segments. Some of these pisolites, however, may have been inorganic. Because the pisolites are almost spherical and could have rolled about readily, they are found in the back-barrier channels as well as on the highs.

The pisolite zone grades landward into even-bedded granular dolomite. Fossils are less numerous and less varied in these bedded deposits than in those nearer the reef. Gastropods, ostracods, pelecypods, and worms are the main types present. These are the forms best qualified to live in restricted waters with sun-heated bottoms. These impoverished faunas have been found as much as 15 miles behind the reef front (Newell, 1939). Dolomitization may be responsible, in part, for the lack of fossils, but casts and molds are as rare as are complete forms. In subsurface specimens the worm tubes are commonly filled with clear or brown anhydrite or ankerite.

Near the boundary of the carbonate and anhydrite zones the dolomites take on a porcelaneous lithographic texture. Anhydrite inclusions become very common and eventually make up more than half the rock. It is probable that most of these carbonates are chemically precipitated primary dolomites, but some may be clastic muds washed in from the near-reef areas.

The back-reef carbonates are interbedded with layers and lenses of sandstone and silt. The Yates sands are best known. These sand members cross the back-barrier mound reefs but normally thin over their crests and thicken in the intervening channels. The channel sands are commonly red, but over the mounds they are gray or white. Pyrite crystals and nodules of pyrite are common and in outcrop appear as small nodules of secondary hematite, pseudomorphs after pyrite. Sands at the lagoonward edge of the carbonate zone are normally cemented with anhydrite. Basinward the sands pinch out at the back edge of the reef wall. It is probable that retreating waves carried much of the loose sand over or through the wall and deposited it in the basin beyond. Coarse, frosted sand grains are scattered through the fine sands. The frosting may be the result of wind action, or it may have been caused by the attack of concentrated brines in the lagoonal areas. Under high magnification the faces of the larger sand grains show minute concussion fracture cones. The thin flakes around these cones would offer little resistance to solution, and they could easily be loosened by the crystallization of brine salts which had seeped in along the cracks. None of the frosting is due to growth of secondary crystal faces. Dolomite, anhydrite, and pyrite are the common cements, but much of the sand is quite friable.

The anhydrite, salt, and near-shore clastic zones of the Capitan differ very little from those of the earlier Permian back-reef deposits, except that the Capitan clastics are mostly fine sand, whereas those of the older formations are mostly shale. The only prominent shale in the back-reef section of the Capitan are numerous lenses of bentonite. Most of the bentonite is a pale-green or gray waxy shale. It was probably widely distributed over the southern Permian Basin, but it is seldom recognized in the anhydrite and salt sections.

Back-reef carbonate rocks can usually be identified by a microscopic examination of the cuttings and cores. Colors

range from white through gray and tan to brown. Anhydrite inclusions are common from the pisolite zone inward. Bedding is regular and uniform, except for the sand beds, which show flow structures, ripple marks, and contemporaneous erosion. The back-reef carbonates grade into the reef wall and are difficult to separate from it, but the highly fossiliferous back-reef and pisolitic zone can be easily recognized in thin sections and even in chips. In cuttings most of the calcite is so coarsely crystalline that it would be regarded as secondary. Coarse porosity is common in many places. Almost all the back-reef carbonates are slightly stained with hydrocarbons. They give off ammonia fumes and are commonly fluorescent. Most of the rock yields organic residues when dissolved in acid. Black and amber asphaltic material occurs in some of the more porous lenses.

It is commonly assumed that dolomitization is the cause of macroporosity in the reef and back-reef beds. This is probably incorrect. Most of the porous rock was made up of shells, pisolites, algae incrustations, calcareous sand, and other fragmental carbonates. The porosity today is only a residuum of the original openings after cementation and other diagenetic processes. In rocks with high original permeability there has been a reorganization of the pore spaces. Some of this may have occurred during dolomitization, but most of it is the result of ground-water circulation. The over-all trend seems to have been a decrease rather than an increase in total pore space. Those dolomites made up of carbonate muds do not have a coarse porosity and appear to be no more porous than undolomitized clastic carbonates of the same grain size. Carlsbad Cavern, the best known of the macropores, is clearly a product of solution and redeposition in a highly porous rock rather than a result of dolomitization (Bretz, 1949).

Fore-reef.—The Capitan Reef is an asymmetrical, flat-topped structure. Beds on the steep front face dip basinward at angles up to 30°, with an average around 20°. Topographic relief on this front varies from about 1,200 to more than 1,800 feet. On the upper slopes just in front of the reef wall and within the zone of photosynthesis the rock is largely a felt of filamentous algae, with numerous interspersed sponges, heads of colonial algae, and other fossils. The filamentous algae apparently swung back and forth and up and down with the waves and helped to break the force of the water. The bedding is rather rudely set off in massive, lumpy layers. Some of this apparent bedding may be secondary. The rock is highly porous. Interbedded with the fossil layers are a few lenses of granular limestone. This upper zone of massed fossils extends some 200–250 feet down the slope.

The second zone on the fore-reef slope comes in below the zone of active photosynthesis. Algae are largely absent, and other fossils are less abundant than near the crest. Much of the rock is made up of boulders and breccia broken off the growing face of the reef. Some of the jumbled masses of breccia look as though they had been emplaced by landslides. Lime muds and coarsely granular limestones form the matrix of the breccias. Mixed with the carbonates are lenses of sand and silt. The steep dipping surfaces on which these lenses rest are irregular, and the outcrops are wavy. Where siliceous material is missing, lenses of lime mud may take its place. Bedding planes are better marked than in the higher beds, but the rock is still very massively

bedded. The beds in this part of the section are made up largely of talus and reef debris. The dip of the bedding planes is very close to the angle of rest. There is little reason to expect that carbonates would be chemically precipitated in waters of this depth, and the crinoids, pelecypods, sponges, and other forms that appear to have grown in place do not make up more than a small fraction of the rock. Thin sections show that recognizable fossil fragments are very abundant (pl. 2, *A*). Tongues of the dark, fine-grained pontic facies rocks extend for as much as 1,000 feet up the slope. Many of the fine-grained beds show extensive crumpling and slumping.

Near the toe of the slope, where the dip begins to decrease, the detritus becomes finer. This fine material forms a basinward-sloping bottom-set fan comparable both in shape and origin with the alluvial fans along the base of a steep mountain front. Because of the fineness and slow rate of accumulation, the bedding planes are thinner and much more distinct than are those higher on the slope. Fingers of bottom-set limestone extend down the slope of the fans for several miles before grading into the pontic clastics. The sections between these fingers grade abruptly into sand near the break in the slope.

The limestone tongues are of different ages along the reef front, and their width also varies greatly. King (1948) gives a very thorough discussion of this interfingering in the vicinity of Guadalupe Peak, where he has named and studied the various tongues. The upper, or Lamar, tongue is the only one that can be recognized completely around the basin.

In the early stages of Capitan Reef growth, the barrier was relatively low and the fore-reef slope correspondingly short. As a result, the same amount of debris would build the talus slope forward much faster than in the later, higher stages. Because the waters were shallower, many more organisms grew on the lower slopes and even on the basin floor in front. Even though the relief was low in the early stages, the reef crest was growing forward as fast as it was able, and the dip of the fore-reef slope was maintained at about 20°.

Normally, carbonates deposited in deep marine waters are calcitic. The algae seem to have contributed a considerable amount of magnesium, and where algae grew on the lower slopes the rock may be dolomitized. The main factor, however, that determines whether the fore-reef beds are limestone or dolomite is whether the detritus from which they were built was derived from the dolomite beds of the reef or back-reef or from the limestones of the upper fore-reef slope.

Only a few wells have encountered the Capitan Reef mass on the fore-reef slope. Some few of these were drilled to test pinch-out possibilities in the sand wedges at the toe of the reef. Sample examinations show fine to coarse detrital limestone, with occasional stringers of dolomite. Sand beds normally come in near the base of the section. All these wells encountered the Capitan at subnormal elevations and because they were far down the slope penetrated a thin section of limestone above the underlying basinal deposits. Several other wells, drilled behind the barrier, have penetrated foreset beds under a capping of back-reef and reef-wall limestones. Most of the rocks are medium to finely granular gray limestone, commonly highly fossiliferous, with a few beds of dolomite. Adjacent wells encounter different types of sections, as might be expected from the steep initial dips.

Basinal facies.—The deep-water equivalents of the Capitan include the pontic beds of the Bell Canyon formation in the Marfa and Delaware basins and the Grayburg beds of the southern Midland Basin and the Sheffield Channel. Shallow-water deposits of the eastern depressions include limestones and dolomites (pl. 2, *B*), some of which are built up into isolated mound reefs. In the deeper areas of the Midland Basin these carbonates give way to sands. In the western basins the main mass of the bottom-set beds consists of flat-lying sandstones and silts. The sands are mostly fine to very fine, but near the edges of the basins some frosted quartz grains comparable to those of the back-reef Yates and Queen formations are present. The percentage of silt increases toward the center of the basins, but the silts are not so important in the Capitan equivalents as they are in the older formations. The only true shales in this part of the section are the bentonites, which are present in several zones.

Along the margins of the basin, wedges of limestone extend out from the toe of the reef to interfinger with the clastics. In the marginal areas these beds are gray or brown, but a short distance basinward they become dark and dense. Most of the rock is limestone, but silica and dolomite are present. The percentages of silt and sand increase basinward until all traces of limestone are lost. The limestones and many of the sands are fossiliferous. Basinward limestones show very little fluorescence; reefward this property is more marked. The upper sands near the center of the basin almost all show oil, and in some areas they are productive. Sections outcropping in the eastern Delaware Mountains and electric logs and sample logs of wells drilled in the center of the basin indicate a remarkable uniformity of the bedding in the truly pontic deposits.

REEF STRUCTURE

The Capitan Reef grew on an inherited structural foundation. Structures contemporaneous with growth are largely either depositional or are the results of differential settling. Some may be a combination of both.

As mentioned above, a cross section of the Capitan Reef sequence is remarkably similar to that of the conventional picture of a delta. Fore-set beds of the reef front advanced across bottom-set beds of the basin floor, while older fore-set beds were being buried by new top-set beds. Because the basin was subsiding faster than bottom-set beds were being laid down and because the reef maintained itself at or near sea level, the only modification of this plan was that the fore-reef slope became higher and higher as the reef advanced.

Back-barrier mound reefs grew where large quantities of water could cross the reef. These mound structures are limited to the portions of the back-reef area that overlie original basins and are missing in the areas that overlie original platforms. The dividing line between original basin and original platform is marked by asymmetrical, down-to-the-basin folds. Off-structure thickening of the beds along the basinward margin of these folds shows that they were growing during Capitan time.

There are numerous small compression structures in the denser, thin-bedded, pure carbonate rocks of the back-reef zone. Most of them are located in the zone where the back-reef beds begin to lose their massiveness. They seem to follow vertical fractures, along which part of the rock has been partially removed by solution. The remaining ledges have been

jammed together to form steep crested buckles, which are commonly somewhat crumpled. The buckled zones are generally 4 or 5 feet wide, and from 5 to perhaps 20 or 30 feet in height. The tops and bottoms are cut off sharply by interbedded bands of silt or perhaps by beds of pisolites. In a few cases buckling is renewed directly above or below the limiting clastic zones. In such cases the buckled zones may extend from the top to the bottom of the canyon wall, but more commonly there is a decided offset or a complete discontinuity. The cause of the buckling is not apparent, but the accompanying deformation is clearly contemporaneous with reef growth. In the field, for the lack of a better name, these features were referred to as "tepee structures."

Several thin sandstone dikes cut the fore-reef and back-reef rocks. Most of the dikes are thin and irregular in outline. The sands are similar to those of the Bell Canyon and Yates formations. Probably the dikes were intruded into fractures formed during reef growth.

POSTREEF STRUCTURE

The Capitan is the youngest of the Guadalupian formations in Texas (Adams *et al.*, 1939). Since the close of the Guadalupian epoch the Capitan and its enclosing formations have been subjected to almost continuous warping and to occasional periods of faulting, folding, and intrusion. In some of these movements the trend of the deformation was determined or localized by the presence of the reef mass or its associated structures. Others were of a regional nature.

At the close of Capitan deposition the whole Delaware Basin was uplifted and tilted eastward. Several additional tilts are recorded in the late Permian, and the western rim of the basin did not again sink below sea level until Cretaceous time. In the Tertiary or later time this western rim was faulted down into the Salt Flat graben. Faulting along the east side of this graben has exposed a cross section of the Capitan Reef in the southern Guadalupe Mountains.

In the Glass Mountains, Chinati Mountains, and other structures on the southeast side of the Marfa Basin and in the Davis, Wiley, and Van Horn mountains on the west, the Cretaceous rocks rest on eroded surfaces of Permian and pre-Permian beds. The same is true in the Las Delicias area of Coahuila. Apparently these areas were uplifted and deeply eroded in the interval between Permian and Cretaceous. A deep Cretaceous and perhaps Jurassic geosyncline seems to have occupied much the same position as did the assumed Permian geosynclinal seaway across north central Mexico, but the area is too deeply buried by these later sediments and by lava flows to permit delineation of the early structures.

The Capitan Reef with its associated heavy-bedded limestone forms a competent strain-resisting mass. Along most of its course this mass is enclosed in more pliable beds. As a result, the reef mass has tended to act as a unit in resisting and in channeling deformation. This is well illustrated along the north edge of the Apache Mountains, where a Tertiary fault parallels the front of the reef for almost the entire length of the range. The early platforms or reef-bordered shelves that furnish the foundation for the Capitan were likewise competent units. In most areas the more massive parts of the Capitan are miles in front of the edges of these old competent foundations. The thin-bedded back-reef beds spanning the space between are much less competent than are the heavy

limestone masses on either side. When stresses acted on the foundation and on the massive Capitan, the deformation was concentrated along this medial zone of weakness.

There is a conspicuous double line of sharp compressional anticlines and synclines in the exposed back-reef area west of Carlsbad. They follow the weaker zone between the back edge of the massive Capitan and the front edge of the old foundation. These folds apparently formed during the post-Capitan tilting of the area. The sharp margins and parallel elongation of the folds set them apart from the rounded mound reef domes on their basinward side.

In spite of the numerous deformational stresses to which it has been subjected, the Capitan Reef is remarkably well preserved, and most of the original structural and sedimentary features can be made out.

DEPTH OF WATER IN THE DELAWARE BASIN AT THE CLOSE OF CAPITAN REEF GROWTH

All the geologists who have worked on the Capitan Reef realize that the pontic deposits of the Delaware Basin were deposited in deep, quiet waters. Several of the workers have attempted to estimate the depth in feet. King (1934, p. 788) first suggested a figure of 500 feet but later (1942, p. 638) increased his estimate to 1,000 feet or more. Adams (1936, p. 789) first suggested a depth of 1,800–2,400 feet but later (1944, p. 1598) reduced it to a depth of 1,200 to 2,000 feet. Kroenlein (1939, p. 1684) estimated a depth of 1,800–2,000 feet for the New Mexico portion of the basin.

The simplest and most accurate method of determining the depth of water in a natural basin is to measure it. In basins of the geologic past, like the Delaware Basin, direct sounding is no longer possible. A substitute method which is reasonably accurate is required.

Commonly used methods involve the use of diagnostic sediments and/or fossils with limited ecological ranges. Shallow-water sediments can generally be recognized as such by their physical characteristics and by the nature of their included structures. For intermediate depths these criteria are inconclusive or, at best, only relatively correct. Fossils are a little more specific than sediments, but in the intermediate ranges they also break down, especially when they are fossils of the distant past. Criteria depending upon temperature, salinity, and light-penetration can be no more exact than the variability of the factors that they measure. Bottom pressure at the time of burial is the only exact physicochemical factor, and at present we have no means of measuring it.

There is, however, a very useful and accurate method that may be applied in special cases. In recently abandoned lakes, where preserved shore lines show little evidence of warping and where the bottom has not been extensively altered by erosion or sedimentation, the original depth of water can be taken as the difference in elevation between the preserved strand line and the basin floor. Accuracy of the measurement depends upon the exactness in determining the original bottom and surface levels. In older buried, warped, or eroded basins the critical data for this type of measurement are seldom available and the method is not commonly used.

One of the places where this method might be used is the Delaware Basin. Here the reef crest and the higher spots of the back-reef area stood at or near sea level. The edge of the reef was separated from the basin bottom by a fore-reef

slope with an average basinward dip in excess of 20°. Distances between points at which sea level and basin floor can be measured is only some 3 or 4 miles. Structural relationships between the top and bottom levels are apparent. There has been post-Capitan tilting, but, when the area was tilted, it was tilted as a block. In the northern end of the Delaware Basin, where the best data for measurement are available, the reef mass rests on the same prereef pontic basement as do the basinal reef equivalents. There is no evidence of differential compaction or subreef gliding. Because the pontic beds in front of the reef are composed of only slightly compressible sands and coarse silts and the reef mass of equally uncompressible limestones, it can be assumed that the vertical distance today between the top and bottom levels is approximately as great as the depth of the Delaware Basin was at the close of Capitan Reef growth. The paleophysiographic method of depth measurement described above seems, in the case of the Delaware Basin, to fit the requirements for sounding by proxy better than any other method so far suggested.

The reef wall formed a ridge, along the edge of the Capitan Reef, that kept most of the back-reef material from being washed out into the basin. It was not absolutely effective, and enough loose material was carried out to form a bottom-set fan along the toe of the fore-reef slope. The surface of this fan was higher at all times than the surface of the true pontic bottom-set beds beyond the limits of the fan. To get an approximately correct figure for the depth of the basin, it is necessary to make the measurements far enough out from the toe of the reef to be near the outer limits of this sedimentary fan but not so far that minor regional warping and other variations would be sufficiently compounded to interfere materially with the accuracy of the measurements.

The whole area along the northern edge of the Delaware Basin has been tilted eastward since the close of Capitan deposition. There are possibilities of error in the measurement of differential elevations along a dip-line projected far enough beyond the reef to reach the edge of the bottom-set fan. To offset this possible error, a strike projection should be measured at right angles to the dip. The fortuitous spacing of wells in the northern Delaware Basin prevents the use of a well-datum as a point for the intersection of these right-angle projections. As a substitute, structure contours on the top of the basinal Bell Canyon formation can be used. In an area west and southwest of Carlsbad, measurements on both projections give a depth between 1,850 and 1,900 feet. These rough approximations agree closely with the assumed depth of 1,800 feet used in this paper.

As a check of this measurement, the thickness of the Castile evaporite, a post-Capitan formation which filled the basin to its rim, is slightly over 1,800 feet in the northern Delaware Basin. Near the center of the basin the thickness of these Castile evaporites is in excess of 2,100 feet (Adams, 1944, p. 1604). A deepening toward the center of the basin would seem logical, but it is not certain that this increased thickness indicates a greater depth of water in Capitan times. Deep-water deposits so far from the margins of the basin may have contained enough compactible material to account for the variation. In the southern end of the Delaware Basin, where subsidence appears to have been slower, the depth of water was probably between 1,200 and

1,500 feet. The measurements given are for the depth of water at the close of Capitan Reef growth. Earlier in the epoch, depths were less.

SUMMARY

The Capitan is a classical example of a large barrier reef, miles wide, thousands of feet thick, and hundreds of miles long. The reef was big enough to control the sedimentation of an extensive epicontinental sea and effective enough as a barrier to cause the precipitation of thick lagoonal evaporites. The area in which the Capitan Reef grew is readily accessible, and the structural and sedimentary record is complete enough for an accurate interpretation of the complex history. The reef is essentially a structure of subsidence and is the last of a series of barrier reefs which surrounded the pontic basins of the Southwest throughout most of Permian time. The growth span of the Capitan itself was long enough for the development of a complete barrier-reef complex. Postdepositional history has been simple. Quick burial to shallow depths preserved most of the reef features and also preserved the associated petroleum which furnishes an economic incentive for reef study. Subsequent deformation and erosion have served more to expose than to destroy the primary features.

Preliminary data from the work of Norman D. Newell and associates on the Capitan Reef will soon become available. Our knowledge of barrier-reef geology, ecology, and growth is limited. The Capitan is one of the most completely formed and best-preserved barrier reefs known. The reef and its associated beds offer unusually favorable opportunities for study. For these reasons it is hoped that the co-operative group research, currently in progress, will be continued until the effects of reef growth on the surrounding sediments are fully understood.

ACKNOWLEDGMENTS.—Ideas and observations presented in this report were secured from the files of the Standard Oil Company of Texas and by extended discussions with most of the geologists who have worked on the problems of Capitan sedimentation and stratigraphy in the last twenty-five years, especially with E. Russell Lloyd. Microphotographs were furnished by William J. Plumley.

REFERENCES CITED

ADAMS, J. E. (1932) Anhydrite and associated inclusions in the Permian limestones of West Texas: Jour. Geology, vol. 40, pp. 30–45.

——— (1936) Oil pool of open reservoir type: Am. Assoc. Petroleum Geologists Bull. 20, pp. 780–796.

——— (1944) Upper Permian Ochoa series of Delaware Basin, West Texas and southeast New Mexico: *ibid.*, Bull. 28, pp. 1596–1625.

——— *et al.* (1939) Standard Permian section of North America: Am. Assoc. Petroleum Geologists Bull. 23, pp. 1673–1681.

BLANCHARD, W. G., and DAVIS, M. J. (1929) Permian stratigraphy and structure of parts of southeastern New Mexico and southwestern Texas: Am. Assoc. Petroleum Geologists Bull. 13, pp. 957–995.

BRETZ, J H. (1949) Carlsbad Cavern and other caves of the Guadalupe block, New Mexico: Jour. Geology, vol. 57, pp. 447–463.

CRANDALL, K. H. (1929) Permian stratigraphy of southeastern New Mexico and adjacent parts of western Texas: Am. Assoc. Petroleum Geologists Bull. 13, pp. 927–944.

DEFORD, R. K., and RIGGS, G. D. (1941) Tansill formation, West Texas and southeastern New Mexico: Am. Assoc. Petroleum Geologists Bull. 25, pp. 1713–1728.

HILLS, J. M. (1942) Rhythm of Permian seas, a paleogeographic study: Am. Assoc. Petroleum Geologists Bull. 26, pp. 217–255.

JOHNSON, J. H. (1942) Permian lime-secreting algae from the Guadalupe Mountains, New Mexico: Geol. Soc. America Bull. 53, pp. 195-226.

KING, P. B. (1934) Permian stratigraphy of Trans-

Pecos Texas: Geol. Soc. America Bull. 45, pp. 697–698.

——— (1942) Permian of West Texas and southeastern New Mexico: Am. Assoc. Petroleum Geologists Bull. 26, pp. 535–763.

——— (1948) Geology of the southern Guadalupe Mountains, Texas: U.S. Geol. Survey Prof. Paper 215.

KING, R. E., *et al.* (1942) Résumé of geology of the south Permian Basin, Texas and New Mexico: Geol. Soc. America Bull. 53, pp. 539–560.

——— (1944) Geology and paleontology of the Permian area, northwest of Las Delicias, southwestern Coahuila, Mexico: Geol. Soc. America Special Paper 52.

KROENLEIN, G. A. (1939) Salt, potash and anhydrite in Castile formation of southeast New Mexico: Am. Assoc. Petroleum Geologists Bull. 23, pp. 1682–1693.

LLOYD, E. R. (1929) Capitan limestone and associated formations in New Mexico: Am. Assoc. Petroleum Geologists Bull. 13, pp. 645–657.

NEWELL, N. D. (1939) Invertebrate faunas of the late Permian Whitehorse sandstone: Geol. Soc. America Bull. 51, pp. 261–366.

PAGE, L. R., and ADAMS, J. E. (1940) Stratigraphy, eastern Midland Basin, Texas: Am. Assoc. Petroleum Geologists Bull. 24, pp. 52–64.

SKINNER, J. W. (1946) Correlation of Permian of West Texas and southeast New Mexico: Am. Assoc. Petroleum Geologists Bull. 30, pp. 1857–1874.

THE LOWER CARBONIFEROUS REEF LIMESTONES OF NORTHERN ENGLAND[1]

GEOFFREY BOND
National Museum of Southern Rhodesia

ABSTRACT

The Lower Carboniferous rocks of northern England show two main facies, the distribution of which is related to the behavior of the floor on which the system rests. Certain areas were relatively stable, and on these the clear-water massif facies was deposited. In other areas the floor subsided more rapidly, and muddy-water basin facies resulted. The reef facies is found at the margins of the massifs in narrow tracts between the two main types and at certain places within the basins. A twofold division is therefore made into basinal reefs and marginal reefs.

The problems connected with the reef limestones include distribution, conditions of deposition, lithology and fauna, tectonics, and erosional history. The classic theories are summarized, and brief accounts are given of typical areas. The basinal reefs seem to have been formed as moundlike accumulations of organic debris, but marginal reefs have several origins. Their tectonic and erosional histories are examined. A feature common to all types is their fauna, which is unlike the normal facies. The difficult problem of correlation between the goniatite zones of the reef limestones and the coral-brachiopod zones of the normal limestones is also considered.

INTRODUCTION

For many years the reef limestones of the Lower Carboniferous in northern England have been the happy hunting ground for fossil collectors and also the subject of lively debates among geologists. The writer feels that a review of these occurrences may be of interest to readers of the special reef issue of the *Journal*. The areas in which the reefs occur will be described briefly, and the relation between the reefs and the regional pattern of Lower Carboniferous sedimentation will be presented. The geological résumé of the various localities is based on descriptions published by others, but the classification of reef limestones is my own.

The generalized succession of the Carboniferous system below the Coal Measures in northern England is shown in table 1.

This is the succession in the area around Cracoe in the Craven district of Yorkshire only, but it illustrates an important feature of Carboniferous stratigraphy in the whole region. It shows that below the Millstone grit the Lower Carboniferous is characterized by two contrasted facies which grade abruptly into each other. The boundary between them coincides approximately with a major tectonic feature, namely, the Craven fault system. Because this fracture system is mainly post-Carboniferous in age, it cannot be directly responsible for the facies change, but it coincides with structural features in the sub-Carboniferous floor, which, as will be shown below, pro-

TABLE 1

GENERALIZED CARBONIFEROUS SUCCESSION BELOW COAL MEASURES IN NORTHERN ENGLAND

Southwest of the Craven Faults	Northwest of the Craven Faults
Millstone grit	Millstone grit
Bowland shales	Yoredale series (rhythmic succession of limestone, shale, and sandstone)
Pendleside limestone and shale	Great Scar limestone
Clitheroe limestone	

[1] Manuscript received November 8, 1949.

foundly affected sedimentation during Lower Carboniferous times. Typical reef limestones are found in a narrow belt between these facies.

These peculiar limestones present many problems. These include their mode of formation, their tectonic and erosional history, their place within the Lower Carboniferous system, their fauna and its environmental significance, and their relation to the general pattern of sedimentation of the regions in which they occur. There is a very large literature dealing with them, and the bibliography given at the end of this paper is by no means complete; it is of necessity selected. Further references will be found in the works cited. Similar rocks with similar problems are found in other countries, such as Belgium, but discussion of these has been omitted for lack of space.

The discussion of these problems was initiated by R. H. Tiddeman (1889), who drew attention to the peculiarities of some conical limestone knolls in the Cracoe district of Yorkshire (pl. 1, *C*, *E*, and figs. 1, 3) lying just south of the north Craven fault. These hills are generally about 200–300 feet high and are more or less circular or oval in ground plan. They are commonly surrounded or partially covered by black shales, or limestones with shales, which contrast markedly with the pure gray limestones of the knolls themselves. Tiddeman noted that the obscurely bedded limestone of the knolls had a peculiar fauna. He was also impressed by the close proximity of such limestones to the powerful Craven fault system, along the line of which the Lower Carboniferous rocks appear to change facies abruptly. His own words give the best account of his views (1892):

> There was lateral pressure going on in the crust of the earth, and the result of this was to pucker up the ocean bed into a series of little folds, and here and there the sea bottom was so raised by gradual movements to sea level. The first result was the breaking up by the waves of the rock so raised. . . . A kind of shoal of angular and more or less rounded fragments resulted, and in those shoals arose colonies of animals to whom such a site was suitable. Slowly these little shoals grew by the continual growth and death of shell-bearing and coral-building animals, and their remains were tossed about and more or less broken by the waves. The resulting debris bound itself together and became a solid limestone rock. . . . Now these singular deposits differ in their arrangement from all other stratified deposits with which we are familiar in ordinary bedded rock-masses. They grew, many of them, to a considerable thickness, some nearly 50 fathoms in height, but instead of extending indefinitely or thinning out gradually, as ordinary strata do, they are very limited in their extent, and end abruptly all round in a steep slope. They form, in fact, low truncated cones, of which the upper part shows rough horizontal bedding, and the steep sides give outward dips all round. . . . It is obvious that if these knoll-reefs commenced on a shoal of fragments broken up by the waves, and grew to 200 or 300 feet in height from the accumulation of the remains of marine animals, the foundation on which they rested must have been sinking, so that the animals could still live in that position as regards the surface which was most suitable to them,

PLATE 1

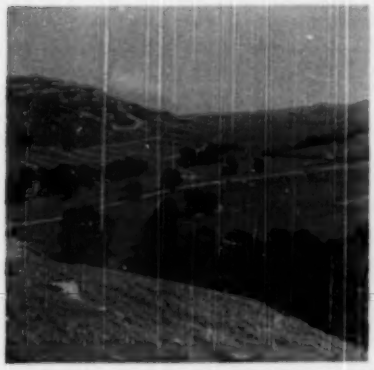
A, The mid-Craven fault scarp at Malham in the Craven reef belt. Namurian shales can be seen burying the scarp in the distance.

B, The railway cutting at Catch All, in the Cracoe section of the Craven reef belt. Namurian shales filling fissures in the surface of D_1 reef limestone.

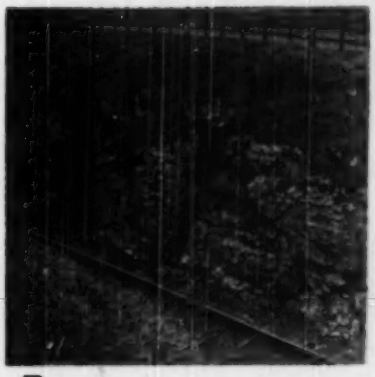
C, Elbolton Knoll, Cracoe district.

D, Fossiliferous reef limestone of D_1 age in the Catch All railway cutting, Cracoe district.

E, Stebden Knoll, which is practically surrounded by Namurian shales. The top of the knoll is level with the base of the Millstone grit, which forms the escarpment behind the knoll. Cracoe district.

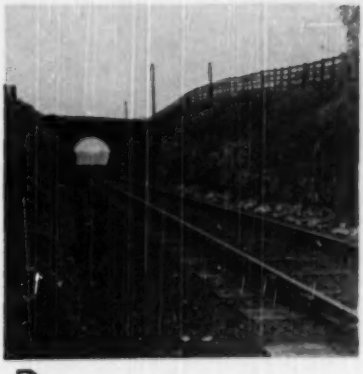


A



B

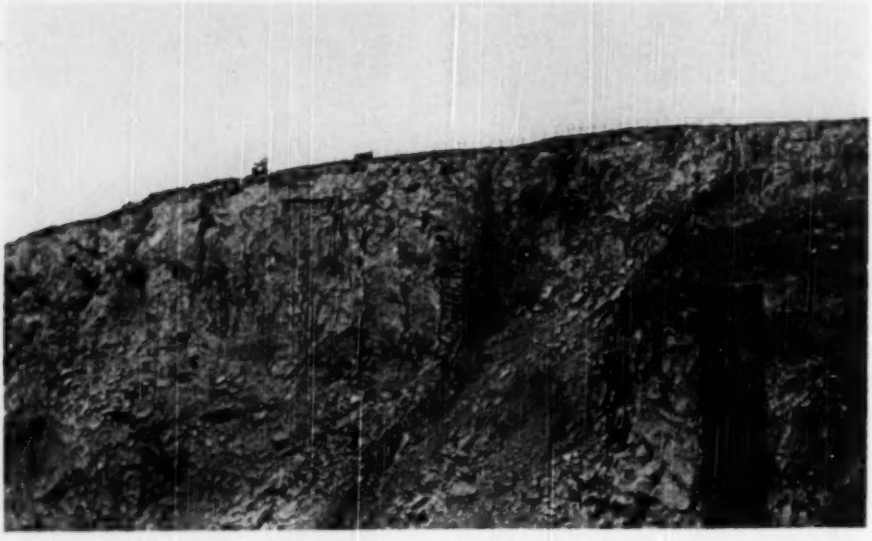
C



D

E

Knolls and reef limestone of Cracoe district



A

B

Quarries in Swinden Knoll, Craven reef belt

that is, within the influence of surface-waves; and when we get reefs like this scattered over wide areas, as is the case, we know that all over the region containing them, the sea bottom was slowly sinking, and, as it sank, the animals kept building up the mounds. Hence, the horizontal bedding throughout the middle of these mounds, and the sloping beds round their edges, which simply represent the material, whether the loosened remains of dead animals or wave-broken masses, big and little, of the reef, which ran down the sides until they came to an angle of rest. . . . The cones grew up as the bottom went down, until either the top was too narrow to give an area of annual growth, or until the rate of sinking was too great for the animals to keep pace with it, and the water consequently became too deep to support such forms under conditions favourable to life.

Such was Tiddeman's view. Marr (1899), however, put forward a totally different explanation. He was more impressed by the association of "knolls" with areas of tectonic disturbance. He wrote (p. 331):

The number of cases where knoll-structure is found in regions of this character, and the fact that no knolls have been discovered in areas where the rocks have not undergone intense plication and fracture is sufficient to show that this is no mere coincidence, but that the existence of the knolls has some definite connection with the disturbed character of the rocks . . . and I proceed to give some account of these detailed structures, premising that the object of this paper is to show that the knolls themselves, with their accompanying breccias, are directly due to the lateral pressure which has affected the district.

These were the "classic" views on the origin of our limestone knolls. Tiddeman emphasized the faunal aspect, Marr the tectonic. Wilmore (1910) struck a much fairer balance between the opposing views, but his work was done before adequate zoning of adjacent areas, without which it was impossible to relate the reef limestones to the regional pattern of sedimentation.

Since the days of Marr and Tiddeman, reef limestones have been described in detail from many other places in the Lower Carboniferous rocks of northern England, Ireland, and the Isle of Man. The distribution of the chief areas is shown in figure 1.

In the early days, work was hampered by lack of adequate zonal division of the system, without which it was difficult to disentangle facies changes or even to distinguish between reef limestones of different ages, although Tiddeman knew that the Clitheroe reefs were on a lower horizon than those of the Craven reef belt.

These deficiencies have been made good in recent years by the detailed zonal work of Garwood and Goodyear (1924), Hudson (1930, 1931, 1935), Hudson and his coworkers (1926, 1935, 1938, 1944), and Parkinson (1926, 1935*a*, *b*, 1936, 1943, 1946, 1949). This has made it possible to assign reef faunas to their proper stratigraphic position, to show the relation of the reefs to the regional sedimentation, and hence to demonstrate the control which the sub-Carboniferous floor had on the distribution of the principal facies types. Zonal work makes possible determination of the relation of the overlying shales, and shales with limestones, to the reef sediments and

PLATE 2

A, The southwestern quarry on Swinden Knoll, Cracoe section of the Craven reef belt. The face is composed entirely of porcelaneous unbedded reef limestone of lower D_1 age.

B, The northeastern quarry on Swinden Knoll, Cracoe section of the Craven reef belt. The lower part of the face is composed of porcelaneous unbedded reef, but fossiliferous grain limestones with well-marked bedding come in in the upper part of the section. The dip is tectonic, not original.

definition of the nature of the contact and the extent of the unconformity which it represents.

It is now possible to test the theories of depositional origin inasmuch as the extent of tectonic disturbance has been ascertained. The result has been a modification of views and a clearer understanding of the picture of Lower Carboniferous sedimentation in general and of the reef limestones in particular.

Garwood and Goodyear (1924), in their work on the Lower Carboniferous of northern England in the region north of the Craven faults, used the same letters for zones that Vaughan used in the standard Avon Gorge succession in southwestern England. This sequence is given in table 2 and has been used as the standard for northern England. This succession unfortunately is based upon the fauna of a clear-water pure limestone facies, and can only be applied in the area north of the Craven faults, where the rocks are very little disturbed and are mainly horizontal. South of these faults the lithology is quite different. Instead of the clear-water limestones, there is a much thicker facies consisting of shales and impure limestones, with a very different fauna. Correlation between the two facies has been difficult until recently but was put on a satisfactory basis by Hudson and Mitchell (1935), and the same zonal letters can be applied to both. The two facies are referred to as the "block" or "massif" facies and the "basin" (or "muddy-water") facies, respectively. These two facies are found throughout the Lower Carboniferous of

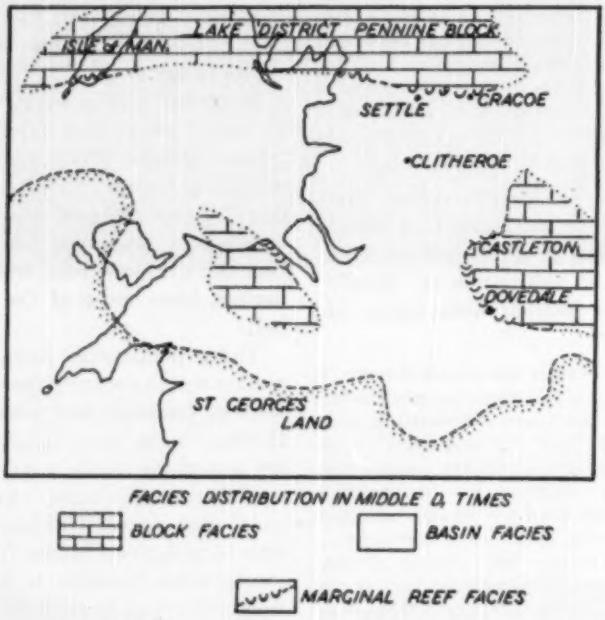


FIG. 1.—Northern England, showing the chief areas of marginal reef limestone and the general relationship of basin and massif areas of sedimentation.

northern England, and their distribution depended on the nature and behavior of the sub-Carboniferous floor. The floor consisted of a number of relatively stable massifs, or blocks, of older rocks, generally tightly packed Lower Paleozoic sediments, with intervening areas which were more subject to subsidence. These rigid blocks formed areas on which massif-facies sediments of clear-water limestone were laid down, and they were not generally submerged until comparatively late in Lower Carboniferous time. In the basins sedimentation began at the very

TABLE 2

SUCCESSION OF THE LOWER CARBONIFEROUS NORTH OF THE MIDDLE CRAVEN FAULT*

Zone	Formation		Thickness (Feet)
D_3	*Orionastraea* band *Lonsdaleia floriformis* beds *Girvanella* band	Yoredale series	100
D_1	Fragmental limestone *Cyrtina septosa* band Fragmental limestone Porcelaneous bed		300
S_2	*Nematophyllum minus* beds *Cyrtina carbonaria* band		200
S_1	Gastropod beds		100
C	*Michelinia grandis* beds		50
	(Great unconformity) Ordovician, Silurian, and Ingletonian—forming the floor of the rigid block		

* After Garwood and Goodyear (1924).

base of the system and resulted in much shale and shaly limestone. Most of the reef limestones are found at the margins of the blocks. Hudson and Cotton (1944) have reconstructed the paleogeography of C_2 times, and their map, from which figure 2 has been modified, shows the distribution of these rigid blocks.

In many cases, the reef facies lies at the margin between block and basin; there are cases, however, where reef limestones occur in areas which are otherwise occupied by rocks of basin facies. This will be referred to again, after a description and discussion of typical areas in more detail.

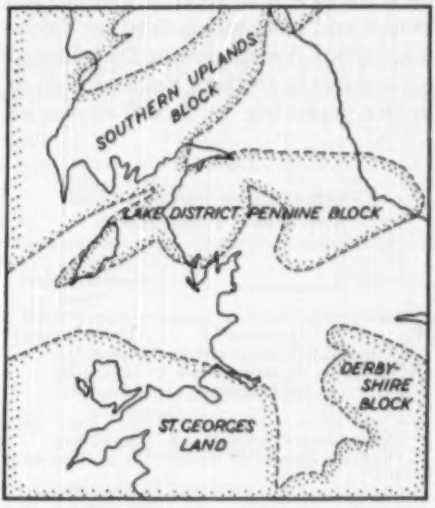


FIG. 2.—The paleogeography of northern England in C_2 times, showing the distribution of resistant blocks of ancient sediments (after Hudson and Cotton, 1944).

Because the fauna of the reef facies does not, as a rule, contain an abundance of the forms on which the standard sequence is based, a separate zonal scheme has grown up for them, based upon the goniatites, which are practically restricted to these abnormal sediments, although they have been recorded in basin facies. The correlation between these two zonal successions has been a difficult and controversial matter, which is by no means finally settled. The goniatite scale is given in table 3 (after Hudson and Cotton, 1944), together with the correlation suggested by the writer's own

work in the Craven reef belt. An alternative correlation will be given in describing the Castleton district in Derbyshire.

DESCRIPTION OF SOME TYPICAL AREAS OF REEF LIMESTONES

THE C-ZONE REEFS

The tendency for reef limestones to form knolls and to occur in well-marked groups and belts was noticed by Tiddeman (1889). A map showing the principal occurrences in northern England is given in this paper (fig. 1). A well-developed group occurs in the Clitheroe-Slaidburn district and has been described in detail in various papers by Parkinson (1926, 1935*a*, *b*, 1936). Except for the reef limestones, the rocks are all of basin facies. The general succession in this area is given in table 4.

TABLE 3

CORRELATION OF GONIATITE AND CORAL-BRACHIOPOD ZONES

Goniatite Zones			Coral-Brachiopod Zones
P_{1a}		*G. crenistria**	(Unconformity)
B	Upper B_2†	*B. delicatum**	Upper D_1
	Lower B_2	*B. vesiculiferum*	Middle D_1
	B_1	*B. holderense*	Lower D_1

* Abbreviations: *G.* = *Goniatites*; *B.* = *Beyrichoceras*.

† B_2 zone is characterised throughout by *Goniatites maximus* (*s.l.*).

The reef limestones form knolls in the Clitheroe area at two distinct horizons. Parkinson's map shows that there is a well-marked anticline trending northeast-southwest on the southern limb of which the knolls occur. The lower series he named the "Coplow knoll series" after a well-known hill, and he assigned it to the C_1 zone. The upper series he named the "Salt Hill–Worsaw knoll series," and he assigned this to the C_2 zone. In both series knolls, or isolated limestone hills, some of them including as much as 500 feet of reef limestone, are common.

Quaquaversal dips are not conspicuous on Parkinson's map, except in the case of Gerna Knoll. However, the regional dip of tectonic origin may have tended to obscure them. Parkinson, like Tiddeman, believed that these knolls formed as distinct mounds of accumulation on the sea floor. They contain a typical reef fauna. The inequalities thus produced were later concealed by the deposition of the Worston shales (shales and impure limestones of S_{1-2} age).

Parkinson has also described a similar succession from the Slaidburn district, a few miles north of Clitheroe. In this district an anticline trending northeast-southwest exposes Lower Carboniferous rocks down to the bottom of the C zone. There are also two strata, C_1 and C_2, in which reef limestones occur. In the two most prominent hills, Great and Little Dunnow, the dips are more or less quaquaversal. These knolls contain a reef fauna and are regarded by Parkinson as conforming to the Tiddeman type.

TABLE 4

SUCCESSION IN THE CLITHEROE-SLAIDBURN DISTRICT

Zone	Formation		Thickness (Feet)
	Namurian shales (Unconformity)		
$D_1 = B_2$, B_1	Pendleside limestone		500
S	Worston shale series		1,200
C_2	Salt Hill series with reef limestone	Clitheroe limestone	2,500
C_1	Coplow series with reef limestone		
Z-C_1	Chatburn limestone		

The reef limestones of the Clitheroe

and Slaidburn areas are marked by obscure bedding and a light-gray to white color. There are sporadic patches of reef tufa and much calcite mudstone. Breccias occur, and in places the rocks are highly crinoidal, in which case they are well bedded.

Parkinson has recently (1949) described reef limestones of C age in the Dovedale area of Derbyshire. Knoll topography is not particularly well developed, but the lithology and fauna are quite typical.

THE D-ZONE REEFS OF DERBYSHIRE

An area which has been the subject of a good deal of discussion recently lies around Castleton in Derbyshire (fig. 1). It has been described in detail by Shirley and Horsfield (1940) and by Parkinson (1943, 1946), who differ somewhat in their correlations and their evaluation of an unconformity in the succession. The area is characterized by a north-facing scarp of more or less horizontal limestones at the margin of the Derbyshire massif or rigid block. The limestones belong to the D_1 and lowest part of the D_2 zones. The scarp was probably produced by a fault with a downthrow to the north, formed in late Lower Carboniferous times. Reef-limestone facies rests on this scarp slope, very much as talus lies on the slope of a modern hillside. This facies probably accumulated as a kind of submarine scree, with initial dips up to 45°. Thus there is a pronounced discordance between the bedded-limestone core and the talus material lying on it. The talus was later covered unconformably by Namurian shales.

Shirley and Horsfield believe that the dip of the reef limestone is always away from the surface of the bedded limestone and that on approaching the flat surface at the top of the slopes the dips in many cases tend to flatten. At the back of the cliff, on the horizontal surface, flat-lying reef limestone may be observed. The successions given by Shirley and Horsfield are given in tables 5 and 6 as the clearest way of indicating their difference of opinion. From these two successions it can be seen that the main point of difference between the authors cited concerns the relation of the goniatite zone of B_2 to the coral-brachiopod beds.

TABLE 5

SUCCESSION IN THE CASTLETON DISTRICT*

Zone	Formation	Thickness (Feet)
	Namurian shales	
	(Unconformity)	
P	Nunlow limestone	0–160
	Calcareous tuff	0–70
B_2	Castleton limestone of reef facies	Variable
	(Unconformity)	
D_2	Gray limestone, coral-brachiopod fauna	~180
	Upper lava	0–20
	Gray crinoidal limestone	0–70
D_1	Millersdale limestone	80
	Lower lava	0–65
	Chee Tor limestone with *Cyrtina septosa* band near top	210
?S_2	Dark limestone	

* After Shirley and Horsfield (1940).

According to Shirley and Horsfield (1940), B_2 lies unconformably upon D_2, whereas Parkinson regards B_2 as the equivalent of the upper part of D_1 because of rapid lateral change of facies. This correlation agrees with that in the Craven reef belt and that in the Pendle district, which lie some 50–60 miles north of Castleton. Both interpretations cannot be correct. At present, it is not possible to disprove either. Leaving aside the

interpretation of structural relationships, a fundamental point may be at issue. It is not possible to map any bed between the two areas, inasmuch as the outcrop is discontinuous. Correlation must be paleontological. Which zonal scheme, that based on goniatites or that based on coral-brachiopod, is the more reliable? Both are based upon faunas which indicate a particular facies and are, therefore, not entirely satisfactory zonal fossils. The coral-brachiopod fauna consists of forms with comparatively simple skeletons that evolved rather slowly. The coral-brachiopod zone includes a key bed, namely, the distinctive limestone with *Cyrtina septosa*. This fossil is restricted to this bed. In Derbyshire there are in some places two such beds, but in Yorkshire only one is common. The goniatites are even more restricted in facies than are the corals and brachiopods, but their skeletons are more complex and their evolution appears to have been relatively rapid. It must be left to future work to resolve these difficulties of correlation.

Whatever the outcome of the matter, Shirley and Horsfield have drawn attention to some very interesting structural features of the Castleton area. Their study of the dips in the reef limestones, coupled with the configuration of the present surface, suggests to them that some parts of the actual coastline of pre-Namurian times, with narrow gorgelike inlets and cliffs, can be accurately mapped, the relief exposed amounting to about 500 feet. They have also found what they interpret as Neptunian dikes of limestone debris in fissures in older limestones.

Typical knolls are absent in the Castleton district. The structures of this area

TABLE 6

SUCCESSION IN THE CASTLETON DISTRICT*

Goniatite Zones	Formation and Characteristic Fossils		Coral-Brachiopod Zones	Formation and Characteristic Fossils	
E_2-R	Edale shales			Edale shales	
	(Unconformity)			(Unconformity)	
Upper P_1	Nunlow beds; *G.* cf. *spirifer*†				
Lower P_1	*G. striatus*		D_2		
	G. crenistria			*Lonsdaleia floriformis*	
	Palaeosmilia regia	Castleton limestone		*Palaeosmilia regia*	
	B. delicatum†			*Dibunophyllum bourtonense* + *Palaeosmilia murchisoni*	Millers Dale beds
B_2	Beach bed			Tuff-bearing conglomerate	
	Cyrtina septosa band		D_1	*Cyrtina septosa* band	
	B. vesiculiferum			Cave Dale lava	
	B. aff. *castletonense*			Chee Tor beds	

* After Parkinson (1946). Parkinson gives two successions, which he claims are lateral equivalents, passing into each other by rapid facies change.

† Abbreviations: *G.* = *Goniatites*; *B.* = *Beyrichoceras*.

are incompatible with the mode of accumulation envisaged by Tiddeman. There was very little folding, although some faulting occurred in Lower Carboniferous times to form the scarp. Yet some of the limestones contain a typical reef fauna.

Parkinson (1949) has recently described reef limestones of D age from the Settle and Cracoe. It was probably once continuous with a similar belt in the southern part of the Isle of Man and perhaps extended farther west into Ireland, where similar rocks were described by Turner (1938). The reef sediments are exposed in three areas now separated as a result of tectonics and erosion between Settle and an eastern cover of younger

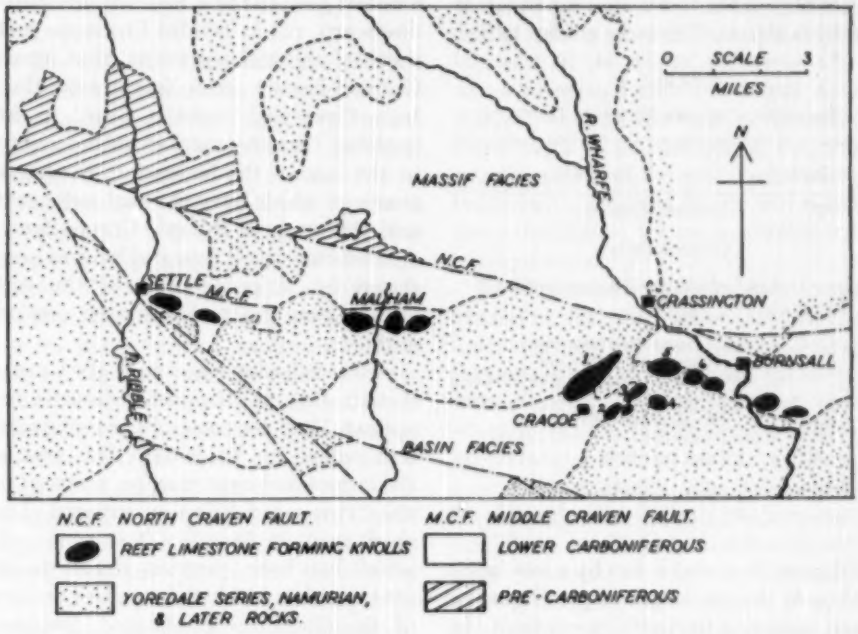


FIG. 3.—The Craven reef belt, showing the division into three parts: the Settle-Scaleber, Malham, and Cracoe sections; and the overlap of the Namurian on the scarp of the mid-Craven fault.

Dovedale area farther south in Derbyshire. The reefs are marginal to the massif facies of the Derbyshire block. He again finds evidence that the B_2 and D_1 zones are correlative. His work has been published only in abstract form.

THE D-ZONE REEFS OF YORKSHIRE

The Craven reef belt (see figs. 1 and 3) is an extensive, though narrow, tract trending more or less east-west through strata. These are the Settle-Scaleber, Malham, and Cracoe sections (fig. 3), all of which have been described by Hudson (1930, 1931, 1938, 1944). The latest description of the Cracoe section is by the writer (1950). Although the three sections are similar, they are different enough to warrant brief descriptions.

The Scaleber area (fig. 3) is the most westerly. The rocks are exposed in a south-facing scarp at the edge of the mas-

sif facies. At the base of the section there are normal well-bedded limestones of S_2 age. These pass upward into reef limestones of D_1 age about 250 feet thick. These are overlain by bedded limestone of upper D_1 and D_2 age. The reef limestones are mainly pale-colored, porcelaneous, calcite mudstones, which are very poorly bedded and very sparsely fossiliferous. A reef fauna occurs at various places in the formation. Knoll topography is absent. The rocks are not folded, although they are broken by a few faults. Some of the faults are of pre-Namurian age, including the mid-Craven fault (fig. 3). Considerable erosion took place in pre-Namurian times, and the overlying Bowland shales rest with marked unconformity on the limestone beds, burying the mid-Craven fault scarp.

TABLE 7

SUCCESSION IN THE SWINDEN DISTRICT

Zone	Formation	
P_1	Namurian shales	
	(Unconformity)	
Upper D_1 and Lower D_2	*Michelinia-Emmonsia* beds (50–75 ft.)	
B_2	Granular and tufaceous limestone with *Cyrtina septosa* band (up to 100 ft.)	Elbolton series
Lower D_1	Porcelaneous beds (up to 200 ft.)	Elbolton series
S_2D_1	Loup Scar beds	Elbolton series
S_2	Threaplands limestone	

The Malham section shows some additional features. The reef limestones occur on the knolls of Cawden and Wedber. These hills consist partly of bedded S_2D_1 limestone of massif facies, although the rocks are locally obscurely bedded. They are truncated by an erosion surface facing toward the basin. On this surface there are screelike remnants that have a steep original dip toward the basin and contain an abundant reef fauna. In their relation to the rocks on which they rest, they resemble some of the deposits in the Castleton district of Derbyshire described above. There is some pre-Namurian faulting at right angles to this scarp related to the mid-Craven fault. This faulting is well shown at Malham, where it gave rise to a well-marked fault-line scarp, cut in bedded limestone just north of the reef sediments, that forms the well-known scenic features of Malham Cove and Gordale Scar. As at Scaleber, there is a marked unconformity at the base of the Namurian, the black shales of which bury the reef sediments and the scarp of the mid-Craven fault. A moderate knoll topography was produced by stripping-away of the soft shales from the pre-Namurian erosion surface.

A few miles farther east, in the Cracoe district (fig. 3), the reef sediments reappear. Here are some of the reef knolls first named by Tiddeman. This area is the largest and most complex remnant of the Craven reef belt now exposed. The chief knoll is Swinden (fig. 3, no. 1), which has been quarried almost to its center and in which the standard section of the district is established. Swinden Knoll is about $1\frac{1}{4}$ miles long and 350 feet high. Among the other knolls are Skelterton (fig. 3, no. 2), Butterhaw, Stebden (pl. 1, *E*), Elbolton, and Thorpe Kail (fig. 3, nos. 3–6), and a few lesser ones. This group of hills has a general northeast alignment.

The Swinden section is summarized in table 7.

The Threaplands limestones are the oldest rocks and are exposed only in the lowest part of Skelterton Hill (fig. 3, no. 2). They are dark, earthy limestones of

basin facies which have been assigned to the S zone. The fauna is quite unlike that found in the true reef facies.

The reef limestones proper begin with the Loup Scar beds, at the top of S_2 or at the base of D_1, which form the first member of the Elbolton limestone series. They are dark, crinoidal limestones with fairly abundant fossils, including typical brachiopods of the reef fauna. They are exposed at the base of the face of Swinden quarry.

The crinoidal limestones are overlain by a thick porcelaneous calcite mudstone of pale cream color and splintery fracture (pl. 2). This is the thickest and most widespread member of the succession. The thickness is variable but reaches a maximum of about 200 feet and makes up the greater part of most of the large knolls. This reef-type rock is obscurely bedded, although bedding can generally be made out in good exposures. Fossils are scarce and occur in small pockets. The fossils include typical reef-facies brachiopods. Enough have been obtained to show that the rocks belong to the D zone. The origin of the mudstone is obscure, but it probably represents quiet lagoonal conditions in which calcium carbonate was precipitated by algae or by direct chemical action. An abrupt change takes place at the top of the limestone, and there is probably a nonsequence at this horizon.

The mudstone is overlain by well-bedded granular limestones. Fossils are common (pl. 1, *D*), one small quarry having yielded over 130 species. This limestone is not more than 100 feet thick and is traceable throughout the district. It is an important key bed in working out the tectonics.

A few forms of *Palaeosmilia murchisoni* and *Cyrtina septosa*, of the massif facies, make possible direct correlation with the standard zones (D_1). This bed also contains B_2 goniatites, which fix its age on this time scale, so that a correlation between the two facies is possible. The great majority of the fossils are, however, of no value in correlation but merely indicate the reef environment of deposition. Many of the fossils show stunting due to adverse environment. The appearance of some suggests periods of normal growth alternating with times during which growth was strongly inhibited. On the whole, the great abundance of species and individuals shows that for most of the time life flourished exceedingly. This member of the series was the source of the great quantities of beautifully preserved fossils which have made the district of Cracoe a famous collecting ground.

The limestones were probably formed in shallow, turbulent water. The granular limestones are not true oölites, although they appear to be so in hand specimens. The granules are fragments of shells which have been broken up and rounded by wave and current action on the sea floor. The mollusks and thick-shelled brachiopods also indicate shallow-water conditions. Reef tufa is abundant in some beds. Because it encrusts individual shells and corals and occurs in irregular sheets along the bedding planes, it is primary. It probably indicates shallow waters, subject to temporary emergence.

These fossiliferous beds are followed by well-bedded, heavily crinoidal limestones, with a scanty coral-brachiopod fauna characterized by the corals *Emmonsia parasitica* and *Michelinia* cf. *tenuisepta*, which, although they are facies forms, are locally useful in correlation. These crinoidal limestones are the highest unit which is widespread in the knolls and include the *Girvanella* nodular band, which is taken as the dividing line

between D_1 and D_2 zones of northern England in the massif facies.

Each member of the above section displays marked lateral faunal and lithologic variation normal to the strike of the reef belt but no such variations parallel to it. The porcelaneous limestones grade into the fine clear-water fragmental limestones of the normal block facies. The highly fossiliferous beds lose their fauna and grade into fragmental limestones which include the *Cyrtina septosa* band in the D_1 zone. The *Michelinia-Emmonsia* crinoidal limestones can be traced southward from Cracoe into the basin sediments and northward into massif facies at the base of D_2.

Just east of the River Wharfe near Burnsall (fig. 3), there is a particularly interesting, though small, area of limestones that was described by Anderson (1928). The succession here is somewhat similar to that in the Cracoe area, but the reef facies is less well developed, and correlation of the strata with the standard coral-brachiopod zones is clearer.

The mapping of the Cracoe district has shown that it is unlikely that the knolls were deposited as discrete mounds, as Tiddeman visualized. Neither can they be explained entirely by tectonics, as Marr suggested. This latter hypothesis did not take into account the peculiar faunal characteristics of the knolls.

The Cracoe district lies at the eastern end of an anticline which trends northeast-southwest across the region of basin facies south of the Craven fault system. This region is much folded in contrast to that north of the faults, where the strata are almost horizontal. It was this fact that led Marr (1921) to name the region the "rigid block," and it was shown in the introduction to this paper that the distribution of such rigid blocks (see fig. 2) determined the general pattern of sedimentation throughout northern England. They also seem to have controlled the tectonics. In the Cracoe district, the structures in the limestones are mainly due to faulting and folding which was completed before the deposition of the Namurian Bowland shales began. The folding affected rocks which were under exceptionally light cover and devoid of all shale partings which could act as bedding-plane lubricants. The rocks, therefore, were shattered, the bedding was rendered obscure, and folds of a rather peculiar pattern were produced.

In the Cracoe knolls area, the main anticline (along which there has been movement in post-Carboniferous times) divides into two branches separated by a broad shallow syncline. The northern branch runs through the long Swinden Knoll (fig. 3, no. 1), which owes its shape primarily to this structure. Dips up to 40° are found on its flanks. The southern branch is more complex and consists of a number of *en echelon* anticlines. There are also cross-faults at right angles to the fold axes. These die out eastward, where the series is thickest, and do not cross the northern Craven fault.

Erosion of the folded and faulted limestones produced the rounded topography of the knolls, most of which are separated by faults. The Namurian shales were then deposited upon this uneven erosion surface (pl. 1, *B*), and their removal during the present cycle of erosion has exhumed the little-modified pre-Namurian topography. The extent of the pre-Namurian erosion can be estimated from the overlap of the shales onto various horizons of the limestone from S_2 to D_2. Thus the origin of the knolls as erosional features carved from once-continuous sheets of sediment affected by folding and broken into blocks by faulting, as sug-

gested by Hudson (1931), is supported by the work of the present writer. It is significant that Skelterton Hill (fig. 3, no. 2) consists entirely of bedded limestone of basin facies and yet is indistinguishable in appearance from the knolls which contain reef-facies sediments. On the other side of the north Craven fault there are three knolls described by Garwood and Goodyear (1924) made of typical massif-facies limestones.

THE REEF FAUNA

As pointed out above, reef limestones are most widespread in the D zone, but such limestones also occur in the C, and to some extent in the S, zone. The faunal lists from all British Carboniferous reefs are rather similar. The most comprehensive are given in papers by Parkinson (1926, 1935*b*), from the Clitheroe reefs of C age, and by Garwood and Goodyear (1924), from the D-zone reef of the Craven reef belt, from Settle, Malham, and Cracoe. The lists are long. Garwood and Goodyear give over 200 forms, and Parkinson includes 114 from Clitheroe. A comparison does not show more than 50 species common to the two, but these include forms which are exceedingly common individuals and which are not useful for zonal work because they are long-range facies forms. Species which are of diagnostic value in zoning are few in number and rare as individuals, as though the reef environment was unsuitable and abnormal for the region.

The composition of the fauna is interesting. From the Cracoe knolls, Garwood and Goodyear recorded 2 species of corals, 48 species of brachiopods, 52 species of lamellibranchs, 43 species of gastropods, and 16 species of cephalopods. To these must be added species of trilobites, Foraminifera, bryozoans, ostracods, echinoderms (including one echinoid), algae, and more corals, collected by the writer. This assemblage is notable for the strong molluscan element, the comparative abundance of trilobites, and the relative scarcity of corals. A comparison of the C faunas with those from the D reef was made by Parkinson (1926, p. 207), and it shows 39 species common to Coplow and Salt Hill (C_1 and C_2); 28 species common to Coplow and Cracoe (C_1 and D_1); 48 species common to Salt Hill and Cracoe (C_2 and D_1); and 20 species common to Coplow, Salt Hill, and Cracoe (C_1, C_2, D_1).

Parkinson notes some differences in facies. Thus the rich lamellibranch fauna of Cracoe is missing at Clitheroe, which has a varied crinoid assemblage not found at Cracoe. Lamellibranchs and gastropods are rare in both the basin and massif facies of northern England, but they are abundant in the fossiliferous beds of the reef belts. Cephalopods are sufficiently common to provide a goniatite time scale; *Orthoceras* and nautiloids are also common. Yet cephalopods of any kind are extremely rare outside the reef facies.

On the other hand, the characteristic corals and brachiopods of the nonreef facies are hard to find in the reef limestones. A few individuals do occur, although they tend to be obscured by the abundance of the local facies fauna. Enough have been found in most places to correlate the various reefs with the standard zones. The field relations of the reef facies generally allow its position to be determined relative to nonreef sediments.

Although mollusks are common, the most abundant fossils are brachiopods. In some cases the limestones have earned the title of "brachiopod beds," being made up almost entirely of well-pre-

served shells. These can be extracted from the rubbly matrix, particularly in weathered outcrops. Common genera are *Dialasma*, *Pugnax*, *Productus s.l.* (particularly the pustulose and semireticulate groups), *Schizophoria*, *Orthotetes s.l.*, and a great profusion of spirifers (*Spirifer*, *Brachythyris*, *Martinia*, *Athyris*, *Reticularia*). The abundance of certain species is best appreciated from Tiddeman's own collection, now in the Skipton Museum, close to the localities from which he collected. *Schizophoria* is represented by about 1,500 clean specimens ranging down to individuals $\frac{1}{8}$ inch in length. There are 3,000 *Pugnax* spp., 450 *Dialasma* spp., about 2,000 *Martinia* spp., and 200 *Pustula subelegans*.

In thin sections the granular limestones show abundant Foraminifera, *Endothyra* and *Archaediscus* being the commonest forms.

Corals are not, as a rule, abundant. *Amplexus* is generally said to be the most common genus, but *Syringopora*, *Caninia*, and *Lithostrotion* can generally be found. In the D reefs of Cracoe *Dibunophyllum* is by no means rare, and *Palaeosmilia murchisoni* is locally quite common. The reef genera seem to have been able to establish themselves at any time during the Lower Carboniferous in those places where the environment was suitable. Very few of them have been used as yet for correlation. With so many individuals available, variation is painfully obvious, and no large-scale statistical work has yet been carried out on communities from different beds. The general aspect of the fauna is similar, regardless of the bed from which it comes. Wherever it occurs, it seems to indicate similar conditions of depth, clarity, and turbulence. In the next sections various conditions of deposition leading to this environment will be discussed.

THE ORIGINS OF REEF LIMESTONES AND KNOLL TOPOGRAPHY

From the published descriptions and from personal observations in the field, the writer is convinced that reef limestones have accumulated in several different ways and that no useful purpose can be served by trying to explain them all in one way, by following either Tiddeman or Marr. Each occurrence should be treated on its merits, without bias, and described in terms which cannot lead to confusion.

The chief considerations are regional setting; sedimentation and resulting sediments; and tectonic and erosional history.

REGIONAL SETTING

This subject was discussed in the introduction to this paper. It was shown that the reef facies occurs chiefly at the margins of the rigid blocks which determined the distribution of massif and basin facies in the Lower Carboniferous. There are also reef limestones which are now completely enclosed in sediments of basin type.

The reefs are, therefore, "marginal" and "basinal" (the terms are Hudson's). Basinal reefs should be treated with some caution, inasmuch as they all occur in the C zone and may perhaps have been marginal at the time of their formation. They may, however, have been due to local reversal of the general subsidence of the basin areas. Hudson (1949) has shown that, in the Malham region of Yorkshire, the reef facies migrated toward the massif as time progressed; the same may prove to have been the case elsewhere. No further subdivision of basinal reefs is here attempted.

SEDIMENTATION

The use of the term "reef" in connection with these sediments has been ques-

tioned. A dictionary definition of reef gives: "a narrow ridge or chain of rocks, shingle, or sand, lying at or near the surface of the water." There is abundant evidence from these rocks that they were formed in shallow water and that emergence took place both during the accumulation of the deposits and after their formation. The marginal reefs certainly formed long, narrow shoals between block and basin, and the term "reef" seems quite justifiable. They were not quite like fringing reefs or barrier reefs, nor were they even comparable to modern coral reefs or *chapeiros*. At least four types of original structures are recognizable among the reefs of northern England. These are reef-knolls in the strict Tiddeman sense, apron reefs, flat reefs, and unbedded reefs.

Reef-knolls.—The term "reef-knoll" (or "knoll-reef") should only be used to describe cases where accumulation took place in the way Tiddeman described. It is more or less synonymous with "bioherm" of Cumings (1932). He defined this as "consisting of any dome-like, mound-like, lens-like, or otherwise circumscribed mass, built exclusively or mainly by sedentary organisms such as corals, stromatoporoids, algae, brachiopods, molluscs, crinoids, etc., and enclosed in normal rock of different lithological character." This term has not been widely adopted in England, and "reef-knoll," which has considerable priority, continues to be applied. Unfortunately, it has been indiscriminately used to describe features which certainly do not conform to the strict definition. Parkinson regards the C-zone reef of Clitheroe and Dovedale as conforming to this type.

Some very perfect examples of true reef-knolls have recently been described by W. W. Black (1949). They occur just north of Grassington, within sight of Tiddeman's original knolls of Cracoe, in a limestone of the Yoredale series on the rigid block. They are quite small, being only a few feet high and a few yards across the base. They consist of cores of porcelaneous limestone, probably of bryozoan origin, which formed little mounds over which crinoidal limestones show quaquaversal dips. They are now exposed by the removal of some of this brashy limestone. This results in true reef-knoll topography.

Apron reefs.—Hudson has suggested this term for certain features to be seen in the Malham and Cracoe sections of the Craven reef belt, and it is also applicable to part of the succession at Castleton in Derbyshire. In this case a scarp, probably a fault scarp, abruptly terminates the massif series of limestones where they face toward the basin, and a scree, or "apron," of limestone with a reef fauna rests with pronounced discordance across the edges of the older series.

Because the scarps appear to be due to faulting, such scarps do not necessarily imply a long interval of time between the deposition of the rocks which form them and the formation of the apron.

The structure consists of three essential parts. There is the *core* of older rocks which is generally of bedded limestone of massif facies or of a type referred to below as unbedded reef; there is the *erosion surface;* and there is the *apron* of later deposits. This may consist of talus-like material derived from the core, freshly accumulated shell and beach beds, or a combination of both. The dips of the apron are original, not tectonic, and may be as steep as 40°. They follow the underlying limestone surface closely. It may be of interest to note that this

mode of sedimentation was stated as long ago as 70 years by Mojsisovics (1879), who gave a diagram of such a structure in the southern Tyrol. His interpretation has since been challenged, but his intention is clear from his diagram.

Flat reefs.—This term is introduced to include large and small sheets of reef limestone, their origin being indicated by their contained fauna. They were laid down horizontally, and in the marginal belts they may have a long extent parallel to the edge of the block, but only a narrow width. The term is not quite synonymous with Cumings' "biostrome," which he defined as "bedded structures such as shell beds, crinoid beds, coral beds, etc., consisting of and built mainly by sedentary organisms and not swelling into mound-like or lens-like forms." There are beds in the reef limestones of northern England which are rich enough in shells to be termed biostromes, but there are others, such as the granular limestones of Cracoe, which have many fossils but cannot be said to have been built mainly by sedentary organisms. They are merely fossiliferous granular limestones of the reef facies, originally laid down more or less horizontally. Their presents dips are tectonic, not original, and "flat reef" seems to be descriptive of their origin. They pass laterally into the fine, fragmental limestones of the massif facies.

Unbedded reef limestone.—Unbedded reef limestone is widespread. Much of the limestone of the Craven reef belt is of this type. This rock is an obscurely bedded porcelaneous calcitic mudstone. Ill-defined planes which suggest bedding are common. Well-bedded limestone may grade upward into mudstone, above which bedded limestones may reappear. Fossils are generally extremely scarce. Those that occur are in many cases clustered together in pockets and include typical reef-facies forms.

Laterally this type may grade abruptly into limestones of normal massif facies. The basinward aspect is not clear because the rock is usually exposed only in scarps facing the basin. The lack of fossils cannot be attributed to dolomitization because the limestone is almost devoid of dolomite. It is exceptionally pure, and leaves practically no residue after solution in cold dilute acids.

TECTONIC AND EROSIONAL HISTORY

The rocks described above have not been greatly disturbed by later earth movements. The Cracoe district has undergone the most severe disturbance, and most of the movement took place in pre-Namurian times. Movement at the edge of the block was responsible for initiating conditions favoring the reef fauna, and also for erosion which generally preceded the deposition of the overlying shaly beds. Erosion of these shales has clearly exposed and little modified the "fossil" topography. Hudson (1931) has shown clearly that much of the present topography of parts of Yorkshire and Derbyshire was controlled by erosion of the D reefs in pre-Namurian times.

Hudson clearly distinguishes between the two major problems involved. First, there is the origin of the peculiar facies found in these limestones and its relation to the adjoining sediments. Second, there is the origin of the topography of the regions in which these peculiar limestones occur. In preparing this review, the writer has tried to keep these problems distinct. Much work remains to be done, but it is hoped that the general picture presented above will not prove too inaccurate, although much of it is, of necessity, an oversimplification. The picture

as we now see it is far more complex than Tiddeman's original conception, but all honor is due his pioneer work, for it was his conception of reefs which laid the foundation for all subsequent work.

The writer expresses his indebtedness to the work of the authors from whose works this review has been drawn and hopes that their views have been accurately, although very briefly, stated.

REFERENCES CITED

ANDERSON, F. W. (1928) The Lower Carboniferous of the Skyreholme anticline, Yorkshire: Geol. Mag., vol. 65, pp. 518–527.

BLACK, W. W. (1950) Paper read to Yorkshire Geol. Soc., January, 1949. (In press.)

BOND, GEOFFREY (1950) The Lower Carboniferous reef limestones of Cracoe, Yorkshire: Geol. Soc. London Quart. Jour., vol. 105, pp. 157–188.

CUMINGS, E. R. (1932) Reefs or bioherms? Geol. Soc. America Bull. 43, pp. 331–352.

GARWOOD, E. J., and GOODYEAR, E. (1924) The Lower Carboniferous succession of the Settle district and along the line of the faults: Geol. Soc. London Quart. Jour., vol. 80, p. 184.

HUDSON, R. G. S. (1930) The Namurian unconformity at Settle and the Carboniferous of the Craven reef belt: Geologists Assoc. Proc., vol. 41, pp. 290–322.

——— (1931) The pre-Namurian knoll topography of Derbyshire and Yorkshire: Leeds Geologists Assoc. Trans., vol. 5, pp. 49–64.

——— (1935) The age of the *Lithostrotion archnoideum* fauna: Leeds Philos. and Lit. Soc., Science Sec., Proc. 1935.

——— (1949) The Carboniferous of the Craven reef belt at Malham, Yorkshire: Geol. Soc. London Proc., Abstracts, no. 1447.

——— and BOOKER, K. M. (1926) The Carboniferous sequence of the Craven lowlands south of the Cracoe reef limestones: Yorkshire Geol. Soc. Proc., vol. 20, pp. 411–438.

——— and COTTON, G. (1944) The Lower Carboniferous boring at Alport, Derbyshire: Yorkshire Geol. Soc. Proc., vol. 25, pp. 254–330.

——— and MITCHELL, G. H. (1935) The Carboniferous geology of the Skipton anticline, Geol. Survey Great Britain Summary of Progress, pt. 2, pp. 1–46.

——— *et al.* (1938) The Harrogate excursion pamphlet: Geologists Assoc. Proc., vol. 49, pp. 293–352.

MARR, J. E. (1899) Limestone knolls of the Craven district and elsewhere: Geol. Soc. London Quart. Jour., vol. 55, pp. 327–358.

——— (1921) The rigidity of North-West Yorkshire: Naturalist, February, 1921, pp. 63–72.

MOJSISOVICS, E. VON (1879) Die Dolomit Riffe von Südtirol und Venetien, Vienna.

PARKINSON, D. (1926) The faunal succession in the Carboniferous limestone and Bowland shales at Clitheroe and Pendle Hill: Geol. Soc. London Quart. Jour., vol. 82, pp. 188–249.

——— (1935*a*) The *Beyrichoceras* zone at Pendle: Yorkshire Geol. Soc. Proc., vol. 23, p. 25.

——— (1935*b*) The geology and topography of the limestone knolls in Bolland (Bowland), Lancs. and Yorks.: Geologists Assoc. Proc., vol. 46, pp. 97–120.

——— (1936) The Carboniferous succession in the Slaidburn district, Yorkshire: Geol. Soc. London Quart. Jour., vol. 92, pp. 294–331.

——— (1943) The age of the reef limestones in the Lower Carboniferous of North Derbyshire: Geol. Mag., vol. 80, pp. 121–131.

——— (1946) The Lower Carboniferous of the Castleton district, Derbyshire: Yorkshire Geol. Soc. Proc., vol. 37, pp. 99–124.

——— (1949) The stratigraphy of the Dovedale area, Derbyshire and Staffordshire: Geol. Soc. London Proc., Abstracts, no. 1451.

SHIRLEY, J., and HORSFIELD, E. L. (1940) The Carboniferous limestone of the Castleton-Bradwell area, N. Derbyshire: Geol. Soc. London Quart. Jour., vol. 96, pp. 271–296.

TIDDEMAN, R. H. (1899) On the formation of reef knolls, British Assoc. Adv. Sci. Rept. (Newcastle), p. 600.

——— (1892) Presidential address to Yorkshire Geological Society: Craven Herald, January 29, 1892.

TURNER, J. S. (1938) The Dublin district vis-à-vis the Craven-Bowland lowlands: Leeds Geol. Assoc. Trans., vol. 5, pt. 4, pp. 208–216.

WILMORE, A. (1910) On the Carboniferous limestone south of the Craven fault (Grassington-Hellifield district): Geol. Soc. London Quart. Jour., vol. 66, pp. 539–585.

RECENT AND PLEISTOCENE CORAL REEFS OF AUSTRALIA[1]

RHODES W. FAIRBRIDGE

University of Western Australia

ABSTRACT

The contemporary and Pleistocene coral reefs of the Australian shelf regions illustrate reef development in a relatively stable epicontinental environment—the most common facies in geology. All reef forms—fringing, barrier, atoll, and platform reefs—and five varieties of coral island are found here. The vertical distribution of corals is controlled mainly by light, a function of turbidity, and the areal distribution mainly by temperature (minimum is about 18° C. average for the coldest month). The shape of coral reefs is generally controlled by wind, wave, and current, although complex forms due to old geomorphologic or structural controls later modified by wind, etc., are also common. Normally, a small reef patch, in a region of a single dominant wind and current, will grow into a horseshoe form, then to an atoll, and will eventually fill up to become a large platform. Coral islands are due either to accumulation of sand or shingle, in which case they tend to have an oval or streamlined shape, or to a Recent eustatic drop in sea level, leading to emerged coral limestones that tend to be eroded into irregular scalloped patterns. "Negroheads" may be formed by such erosion and by jetsam-like accretion on the reef margin.

Reef borings and structural and geomorphologic features show that ecologic, tectonic, and eustatic factors may all play a part in reef development. The reality of Pleistocene-Recent eustatic oscillations is proved on geological and physiographic grounds, regardless of coral reefs. Most of the reefs in the Australian area are believed to have grown up from initial coastal fringing reefs during eustatic lows, but the foundations of many of the deeper reefs (in the Great Barrier Reef and on the outer Sahul Shelf) must have subsided tectonically. Physiographic evidence shows that an antecedent platform (a continental shelf) existed prior to the main development of the reefs.

Sedimentation is extremely rapid in the enclosed Great Barrier Reef lagoon but is slow on the exposed northern and western shelves. In both living and ancient reefs the proportion of actual colonial corals grown *in situ* is extremely small in relation to the enormous quantities of "coralline" sedimentary debris. The ratio of reef structures to areas of "normal" shelf sedimentation, even in the most prolific coral areas, is also very small. The thickness of reef structures in the Great Barrier Reef lagoon has been proved by boring not to exceed 500 feet. Physiographic evidence shows that most of the near-shore reefs are not more than 60 feet thick. Similar evidence shows that in extreme marginal areas (which probably underwent Quaternary subsidence) certain reefs may exceed 1,500 feet in thickness (e.g., on the outer Sahul Shelf).

The Great Barrier Reef borings and those of Funafuti, Bikini, etc., show that tectonic subsidence was fairly rapid in all of them. Concentrations of $MgCO_3$ of about 5–16 per cent are evidence of stillstands; lower percentages are evidence of rapid submergence. Submarine dolomite formation requires reducing conditions—an environment present in the closed atoll ring of Funafuti but absent in the Barrier Reef. Lack of dolomite in Queensland borings is thus not surprising.

INTRODUCTION

Contemporary coral reefs bring to mind primarily oceanic atolls and coral islands of the Pacific and Indian Oceans. The best-known fossil reefs, however, are mostly epicontinental in character and were formed in shelf regions, in paralic basins, and along geosyncline margins. We should first study, therefore, the type of living reef found today on continental shelves and along the mobile island belts. The oceanic types of reefs are exceedingly rare in the stratigraphic record. This, of course, is not surprising because nearly all the sedimentary rocks are geosynclinal or epicontinental in character. Paradoxically, however, the oceanic reefs have been much studied, whereas little attention has been paid to the others. Reefs in the mobile belts have received more attention during the last two decades or so, and the fine series of papers on those of the East Indies, principally by Umbgrove (1928, 1929, 1930*c*, 1947), Molengraaff (1930), and Kuenen (1933), form a satisfactory guide to contemporary reef formation in the unstable geosynclinal belt and its semi-stable marginal shelves.

[1] Manuscript received January 27, 1950.

It is more difficult to find adequate accounts of reefs near the more stable shores of the continents. Perhaps the best-known reefs of this sort are those of Florida, where the Marine Biological Station of the Carnegie Institution at Tortugas Key has played a major part in modern reef research (see especially the work of Vaughan and of Mayor). Elsewhere, on the more stable shores of the Americas, coral reefs are rare, as is also the case along the western coasts of Africa. In contrast, the excellent fringing reefs of eastern Africa and the Red Sea have been known since the days of Darwin (1842). Although studied by Crossland (1902, 1913), Voeltzkow (1907), and others, they would repay further examination.

The interesting reefs on the stable Sunda Shelf (off Borneo and Java) have been studied by the Dutch geologists. But elsewhere about southeastern Asia reefs are generally not prominent and certainly not well known.

The coasts of Australia, though fairly stable, exhibit many reef types. Because the shores extend from the tropical to the cool temperate belt, they span a complete latitude range for reef-building organisms. The eastern and western coasts of the continent present us with environments of contrasting aspect. For these reasons it is thought that a review of the contemporary reefs and their distribution around the Australian continent may be of some use to the stratigrapher and the ecologic paleontologist.

Very little scientific research has been done on the reefs as a whole. The least-known reefs of the world are probably those of northern Australia, which have never been studied by a competent observer. The writer's experience of these is limited to a single air flight to the Northern Territory and along the shores of the Gulf of Carpentaria. Numerous reefs of the Sahul Shelf region were photographed from the air during the late war, and a photointerpretive study of the principal types, which include several varieties of platform reefs and atolls, has been made (Teichert and Fairbridge, 1948). In this connection, the history of exploration (Fairbridge, 1948*b*) and the sedimentation, geology, and structure of the whole northern Australian region has been reviewed (Fairbridge, 1950*d*).

In Western Australia the only coral islands and reefs examined are those of Houtman Abrolhos (Teichert, 1947; Fairbridge, 1948*a*), whereas on the mainland and offshore islands of the southwestern coast the partly coral "sandstone reefs" have recently been studied (Fairbridge, 1950*c;* Teichert, 1950).

In contrast to the western and northern areas, the reefs of Queensland have been extensively studied. The Great Barrier Reef Committee was set up in 1922 by the Queensland branch of the Royal Geographical Society of Australasia and has since published some six volumes on the area. In addition, there are the early studies by J. B. Jukes (1847), Saville-Kent (1893), Agassiz (1898), Andrews (1902), Hedley and Taylor (1908), Mayer (1915, 1918), Davis (1917), Hedley (1926), and others. The most notable work has been published by members of the British Museum Great Barrier Reef Expedition, who spent twelve months on the reef in 1928–1929 (Steers, 1930; Marshall and Orr, 1931; Stephenson, Tandy, and Spender, 1931; Orr and Moorhouse, 1933*a*, *b;* Manton, 1935; Yonge, 1940). Many reefs were visited by the geographers of this expedition (Steers, 1929, 1937, 1938; Spender, 1930).

During the recent Pacific war the author, in conjunction with C. Teichert and F. W. Whitehouse, carried out a general

reconnaissance of the Barrier Reefs by flying boat and by launch, visiting each of the principal reef types in succession and comparing the air photographs with the actual features on the ground. Some time was spent on Low Isles, comparing the latest observations with those made by the British expedition of 1928 (Fairbridge and Teichert, 1947, 1948). Future possibilities of this type of research have been pointed out by Steers (1945).

References to the Australian reef literature are given in the bibliography. Hydrographic charts are essential to a study of Australian reefs. The *Catalogue of Australian Charts and Admiralty Charts of Australian Waters*, issued in 1947 by the Hydrographic Branch of the Royal Australian Navy, Sydney, and the *Australian Pilot*, volumes 3, 4, and 5, are especially useful. General charts and information may also be found in the series of the U.S. Hydrographic Office and the appropriate "Sailing Directions."

The understanding of coral reefs, particularly their environment in stratigraphic terms, requires attack from many sides. It might perhaps be appropriate to lodge a plea for more co-operative excursions and expeditions, with stratigraphers, sedimentary petrologists, paleontologists, ecologists, zoölogists, botanists, chemists, and physical oceanographers. In Australia, in recent years, excursions of this sort have been attempted by the universities of Western Australia, Queensland, and Melbourne to reef or other coastal areas. We can hardly agree with Professor Wood-Jones, who maintained that "the question of the formation of atolls is in reality a zoological one" (1910, p. 134), or with Professor W. M. Davis, who wrote (1934) that really "the origin of reefs is a geological one," or with Professor Stanley Gardiner (discussion in Lenox-Conyngham, 1925) who remarked: "I think we must look particularly to geodesists for further progress in this question of coral reef formation. Botanists, zoologists, geologists, have really come to the end of their tether. . . ." One has only to reflect on the great strides made by biology and geology since 1925 to realize the limitations of such opinions.

REEF TYPES

Almost all the "standard" reef forms are known in this region. The classic barrier reef here extends over 1,200 miles along the Queensland coast. The atoll is rare but is found sporadically off the northwestern coast and on the Queensland shelf. Fringing reefs are well displayed in the north, although they are little known in most areas. Most widespread are those irregular-shaped reefs which do not fit into any standard category: the platform or patch reefs, alternatively bank or hummock reefs, which are found along the coast line from 29° S. latitude in the west to 27°30′ S. in the east. Isolated clusters of reef-building corals are encountered southward as far as 35° S.

Pleistocene and early Recent fossil reefs are found over even wider latitudinal ranges, probably corresponding to warmer ocean temperatures during the interglacial stages and the Recent "Atlantic" climatic optimum. The degree of "elevation" of these fossil reefs corresponds closely with well-recognized stages of eustatically high sea levels. In no part of the continent are there tectonically elevated reefs similar to those of the East Indies and like mobile belts.

Only in one sector are all the reef types represented, and that is in the Great Barrier Reef area of Queensland, the largest reef group in the world. This complex contains far more than simple barrier

reefs (Steers, 1937). The "low wooded" island-reefs, the extensive platform reefs and reef patches, and the more restricted fringing reefs will be dealt with separately. The bulk of the above-mentioned studies have been concentrated on these inner reefs.

FRINGING REEFS

The fringing reef is perhaps the precursor of all other reefs. It grows along the shore under a great variety of conditions imposed by the climate, vegetation, and rock materials of the adjacent land mass. The fringing reefs of the Australian mainland are among the poorest of reef types. The rigorous conditions near the shore, excessive sedimentation, dilution of sea water by fresh, acidic waters from the land—all these play a part in restricting or inhibiting the growth of the reefs. More vigorous fringing reefs are found encircling offshore continental islands, where conditions are more favorable. Relatively little attention has been paid to this reef type, and study of it, therefore, would be most profitable.

The best known of the fringing reefs are those of Queensland, the southernmost of which occur in the rather exceptional environment of Moreton Bay around Peel Island (27°30′ S.). This interesting reef occupies quite a large area, as I have seen from the air, but the inner parts are largely dead and covered with sediments. Dead reef coral is found at Mud Island at the mouth of the Brisbane River (Saville-Kent, 1893; Thomson, 1905). Hedley (1925*d*) reported not less than fifteen living species at Peel Island, despite the fact that the water temperature here drops to 60° F. in winter (see figures in Hedley, 1925*c*). Hedley noted that in Sydney Harbour, where the winter temperature drops to 50° F., there are only two coral species (and certainly no reefs). Other isolated reefs on southern Queensland coast are reported at Flinders Reefs off Cape Moreton (Hedley, 1925*c*), and in Hervey Bay, south of Fraser Island (Saville-Kent, 1893).

In the Great Barrier Reef region the paucity of fringing reefs is striking, in view of the variety and abundance of reefs farther offshore. In the southern half of the region, small fringing reefs may occur off exposed headlands, in some cases in conjunction with wave-eroded rock platforms. There appear to be a few such reefs in the vicinity of Port Curtis (24° S.), and Thomson (1905) reported one near Gladstone at the mouth of the Boyne River, apparently dead and covered with sediment from the river. Small patches of living reef occur again around Keppel Island (Jardine, 1925). They are better developed around Cockermouth Island (map by Steers, 1937), Whitsunday Island (Saville-Kent, 1893; Agassiz, 1898), and some points of the adjacent mainland (see Stanley's sketch map of this area, 1928*a*).

At Stone Island (20° S.) in Port Denison, near Bowen, there was a flourishing reef fringe (photos in Saville-Kent, 1893) which was largely destroyed by the coincidence of a low spring tide and a tremendous tropical rainstorm in the 1918 hurricane, when 19 inches fell in three days (Hedley, 1925*a*; Steers, 1937). There is a complex reef fringe at Holbourne Island, 20 miles northeast of Bowen, with a shingle bank separating the living reef from an inner, slightly elevated, dead reef (Marshall, Richards, and Walkom, 1925).

There seems to be a complete absence of reefs between Bowen and Townsville, whereas, at Magnetic Island, there is a well-developed fringe on the southwest side. A short distance to the north, at Palm Island (18°40′ S.), there are very extensive fringing reefs. The writer has reconnoitered these from the air, and excellent photographs on the reef were taken by Saville-Kent (1893). There is good growth of coral on the outer edge of the reef, but the inner parts are mainly sediment-covered (Agassiz, 1898; Andrews, 1902). Small coral growths occur at Dunk Island, the Barnard Islands (Hedley, 1925*b*), the Frankland Islands, Fitzroy Island (Agassiz, 1898), and Double Island, but little coral is encountered on the mainland shore south of Cairns.

An extensive fringing reef occurs along the mainland shore about 30 miles north of Cairns between Yule Point and Port Douglas, at 16°35′ S. This reef was examined during the war by the writer and Drs. Teichert and Whitehouse. The reef edge lies from 1,000 to 2,000 feet offshore; the reef surface is almost entirely covered with terrigenous muds from the Mowbray River, which empties into the middle

of this reef stretch. An abandoned mouth of the river occurs at the southern part of the reef. Behind the beach is a line of vegetation-fixed sand dunes, fronting an area of beach ridges and swamp connected with this old stream bed.

Living corals are at present confined to the outer reef edge, and behind it there is a dead reef, 1–2 feet below low-water springs, lightly sand-covered, and marked by the radial pattern of seaweed. At 800–1,000 feet from the shore is the face of an inner, extinct reef which forms a terrace nearly 5 feet above low-water springs. It is strewn with broadly rippled banks of sand and mud. There is a spring tidal range of about 10 feet here, during which these ripples are daily built and rearranged. The inner part of this mud-covered reef is in the process of colonization by mangrove. The sandy mud is extremely soft. Its thickness is unknown, as a posthole borer was unsuccessful in penetrating it. As at Low Isles, carbonic acid, liberated at the roots of growing mangroves, eats into the reef, gradually forming large basins in reef rock, which are filled by soft, yielding

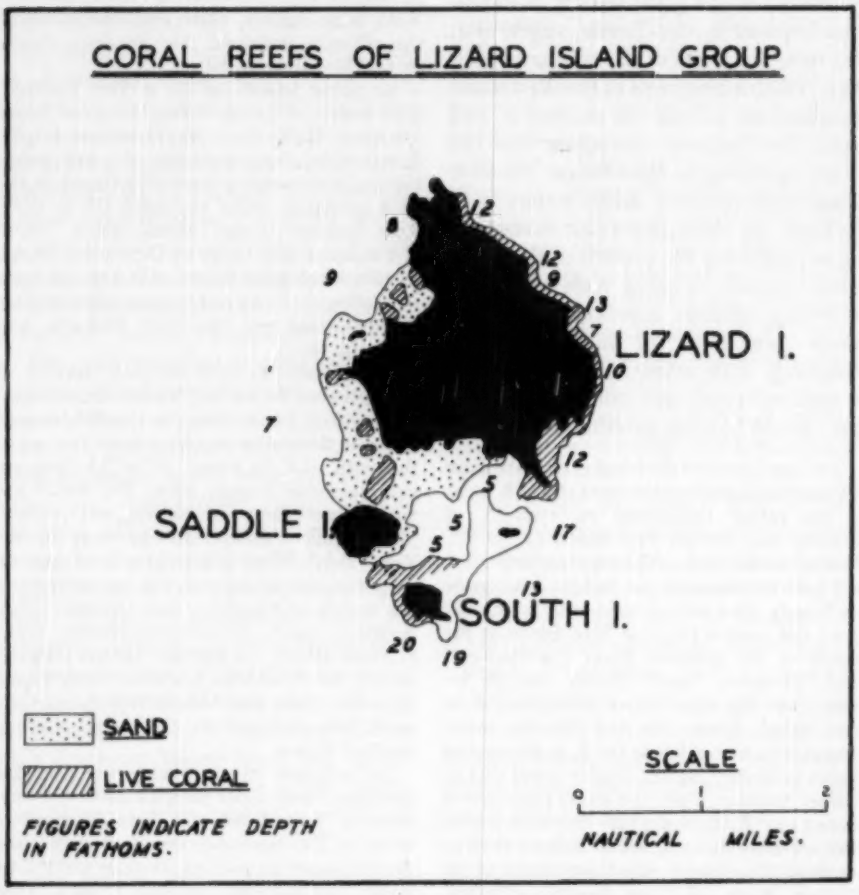


FIG. 1.—Sketch map of the Lizard group, showing typical distribution of fringing reefs around a high "continental" island which rises within the Great Barrier Reef lagoon. Note how the reef is mostly sand-covered on the lee (west) side of the island.

mud (Wharton, 1883; Fryer, 1910, 1911; Spender, 1930; Fairbridge and Teichert, 1947, p. 11; etc.).[2]

Fringing reefs are next found around Snapper Island and along the mainland coast north of Cape Tribulation, where they were seen by the writer both from the air and from the sea, forming an almost continuous belt as far as Archer Point (15°35′ S.). The sandy shores north of Cooktown show only sporadic coral patches.

The Lizard Islands (14°40′ S.), a group of "continental" islands lying nearly 20 miles offshore and not far from the outer barrier, form an interesting comparison with the mainland environment at Yule Point (fig. 1). Stephenson and party described the group (1931, pp. 34, 87), and Teichert, Whitehouse, and the writer visited it in February, 1945 (pl. 1, *A*). The reef corals are much more vigorous and varied here. The interior parts of the reefs are, however, sand-covered. On the south side the reefs join several small granite islands to the main mass and enclose a small lagoon some 5 fathoms deep with a narrow entrance to the east. Coral patches rise from the sandy floor of the lagoon. The level of the reef is mostly between 0 and 2 feet above low-water springs, the spring range being about 8–9 feet.

From Cape Flattery northward to Cape Melville (14°10′ S.), there are almost uninterrupted reefs fringing the mainland coast. Their width varies, as a rule, from 3,000 to 5,000 feet, and most of them appear to be sediment-covered. The continental islands of the Flinders group west of Cape Melville (Richards and Hedley, 1925) are completely fringed, like the Lizard Islands. Past the broad, sandy sweep of Princess Charlotte Bay, the mainland shore is again bounded by a sediment-covered reef. Mangrove is found in all the more protected parts. Similar associations continue almost all along the curious zigzag coast past Cape Direction, Cape Grenville, and so on, right up to Cape York (10°41′ S.).

The continental islands of the Torres Straits, Prince of Wales Island, Horn Island, Thursday Island, Banks Island, etc., are all fringed by similar reefs (photos by Saville-Kent, 1893) and are sediment-covered and partly masked by mangrove. These islands are the last outliers of the eastern Queensland granite and metamorphic rocks, which appear to play a very important role as coral foundations (Haddon, Sollas, and Cole, 1894).

A very different group of islands surrounded by fringing reefs occurs to the east of here, a few miles inside the northern limits of the Outer Barrier reef, facing the Gulf of Papua. They are all volcanic, yet rise near the edge of the continental shelf. They have long attracted attention (Jukes, 1847; Haddon *et al.*, 1894). Among these, Mayer investigated the Murray Islands (9°55′ S., and 144°2′ E.), which consist of three volcanic islets—Maër (Mer), Dowar, and Wyer, surrounded by reef-fringes (Mayer, 1915, 1918*a*). The reefs show a *Lithothamnion* ridge, which tends to impound the water on the reef-flat behind. The maximum width on the southeastern, or windward, side is 2,200 feet. Negroheads, which are so common farther south, are absent. Mayer attributes this to the lack of hurricanes in this area. In consequence, both reef-flat and outer fringe are covered with "one of the most luxuriant coral growths to be found in the Pacific." Parts of the reef-flat nearest the volcanic shore are covered by terrigenous sediment and boulders and in one place by blown sand.

Two other reef-fringed volcanic islands in this area—Darnley and Bramble—with broad, flat reef platforms ½–1 mile in width and devoid of mangrove, have been mapped and described (Jardine, 1928*c*, *d*).

Turning our attention westward, now, to the Gulf of Carpentaria, it is rather astonishing to find this enormous shallow-water platform (averaging 25–35 fathoms) almost completely devoid of coral reefs. Its shores are low and swampy, and large rivers debouch into it. Fringing reefs are encountered locally around rocky islands. At Bentinck Island (17°10′ S.) and Wellesley Island (16°30′ S.), there are discontinuous fringes, and again at Sir Edward Pelew Group (15°40′ S.) (Jensen, 1914; Paradice, 1924). Paradice remarked that muddy waters of the mainland reach well out to the Pelew Islands, becoming dispersed by currents to the north of them. "The coral growing around these outer [northern] islands is more prolific, and much more brilliant in colour, than

[2] The mangrove situation here is analogous to that found in many tropical fringing reefs. On Morotai, in the East Indies, it led to considerable confusion in one of the Allied landings during the war. Great mud-filled potholes in the inner reef engulfed or bogged heavy transport, tractors, and even tanks. Scattered mangrove and the mottled appearance of the inner reef in air photos led the writer to say that trouble might be expected here.

that nearer the mainland." Coral patches occur at Groote Island and at a few rocky points in the Cape Grey area.

At Wessel Island and along the northern coast of Arnhem Land reefs are more extensive. They are in many cases partly buried by sediments and overgrown by mangrove. Long stretches of such reefs fringe the shores of Melville and Bathurst Islands and the long tongue projecting from Cape Van Diemen. The muddy shores of Van Diemen Gulf are also fringed where the current is strong, as in Clarence Strait, and likewise about the harbor and approaches to Port Darwin. These reefs average about 2,000 feet in width, but, where favorable environmental conditions exist, they measure 5,000 and even 10,000 feet across. Their inner parts stand 10–15 feet above low-water springs, exposing an expanse of rippled sand and mud flats. Living coral is restricted to the outermost fringe. These reefs appear sporadically along the shores of Bonaparte Gulf as far as Cape Londonderry, being absent where the Daly, Fitzmaurice, Victoria, and Ord rivers debouch.

Almost nothing is known of these reefs (although a number are indicated on Joubin's map, 1912). The same is true of the coast of the northwest Kimberley, the steepness of which, with its rocky offshore islands, favors some vigorous reef growth. A very great tidal range (up to 30 feet) and constriction of the deep sounds by fringing reefs, especially near the mouths, creates a tremendous tidal bore, which makes examination and exploration of the region very difficult. Along the very steep shores, there are local narrow fringes. In the Buccaneer Archipelago the fringes are rarely over 2,000 feet across. They are broader, however, near the entrance to King Sound.

A fringing reef at the northern end of Troughton Island (13°40′ S., 126°10′ E.) was described by Basset-Smith (1899). In front of the sand beach at this place is a raised bench, beyond which is the reef-flat of coral shingle, half a mile wide. Living coral is absent except in tide pools. On the reef edge there are negroheads and a fringe of rich coral growth, largely below low-tide level. Referring to the northwestern coast in general, Basset-Smith adds: "Another peculiarity of this region was the great turbidity of the water near the coast and the larger amount of slimy mud deposited on the flats, which as fringing reefs were everywhere present, their seaward edge being marked by isolated blackened and elevated masses of coral-rock."

Southwest of Cape Leveque (16°20′ S.) the reefs are few and small. Narrow reef fringes are found between here and Broome and around Lagrange Bay but are interrupted by long stretches of sandy coast line without river drainage, as the interior is semiarid to arid.

Fringing reefs of normal size occur in the Amphinome Shoals area and from Port Hedland (20°19′ S., 118°35′ E.) westward along the coast, around Depuch Island, Dampier Archipelago, and the immense number of little islands offshore between here and North-West Cape, as well as Barrow and the Monte Bello Islands. The mainland reefs are largely sediment and mangrove-covered, those offshore on the rocky islets are vigorous and extensive. The reefs of this 350-mile stretch of coast have not been scientifically studied.

Around North-West Cape there are fringing reefs up to 5,000 feet across, except in Exmouth Gulf, where there seems to be too much sediment for reef growth. From North-West Cape to Cape Leeuwin, over 1,000 miles, and on the

PLATE 1

A, Lizard Island (14°40′ S., 144°27′ E.), oblique air view, looking east, showing fringing reefs around a "continental" island and a shallow lagoon.

B, Batt Reef (16°25′ S., 145°45′ E.), oblique air view, looking west-northwest, showing smooth, curving windward reef edge (rather similar to outer-barrier margin) passing to a radially patterned "trickle zone," and to the interior to a central sand-flat (believed to be an infilled lagoon). (R.A.A.F. official photographs, not to be reproduced without permission.)

PLATE 2

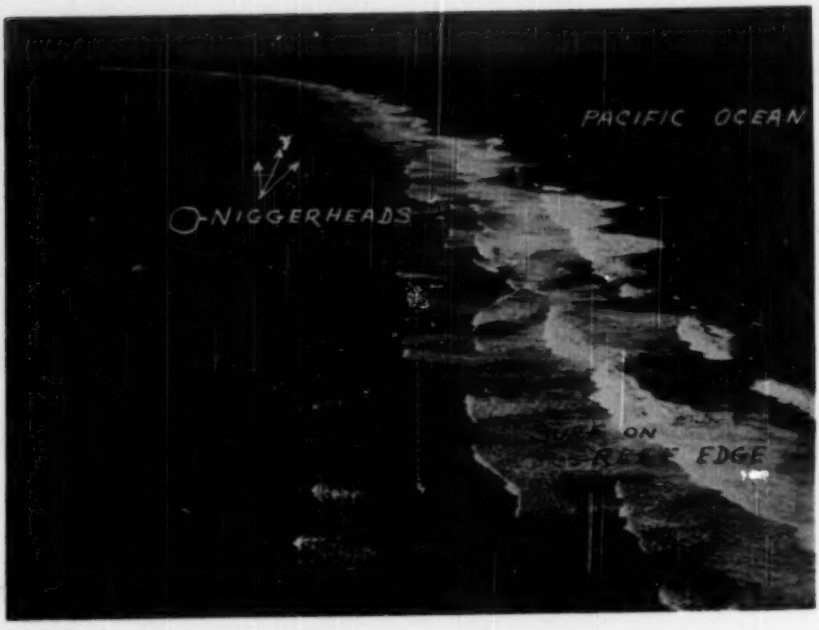
A, Outer-barrier reef, Queensland (Ribbon Reef, 14°50′ S.), oblique air photograph, showing typical smoothly curving outer margin. At high tide, water rushes over reef crest, which is punctuated here and there by negroheads.

B, Outer-barrier reef as in *A*, frontal view, looking west, showing zonal character of reef surface. (R.A.A.F. official photographs, not to be reproduced without permission.)

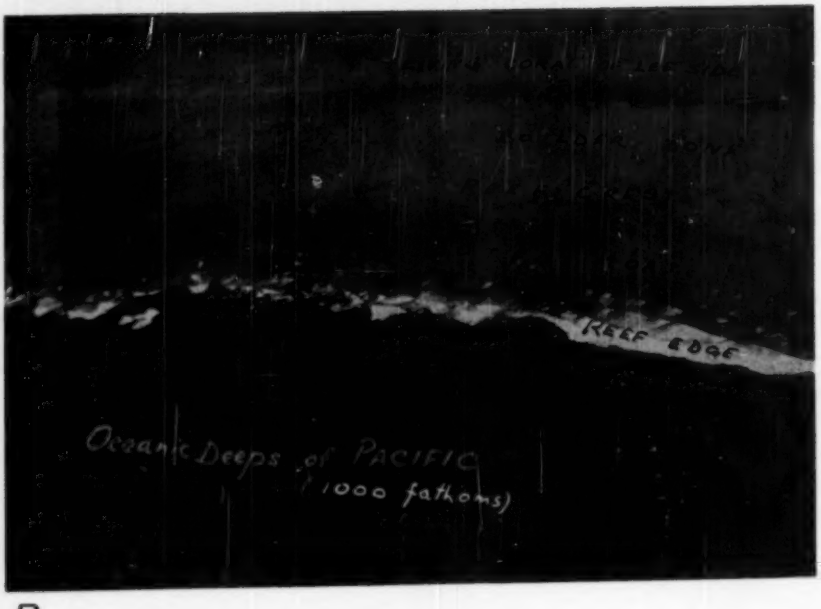
A

B

Lizard Island (A) and Batt Reef (B)



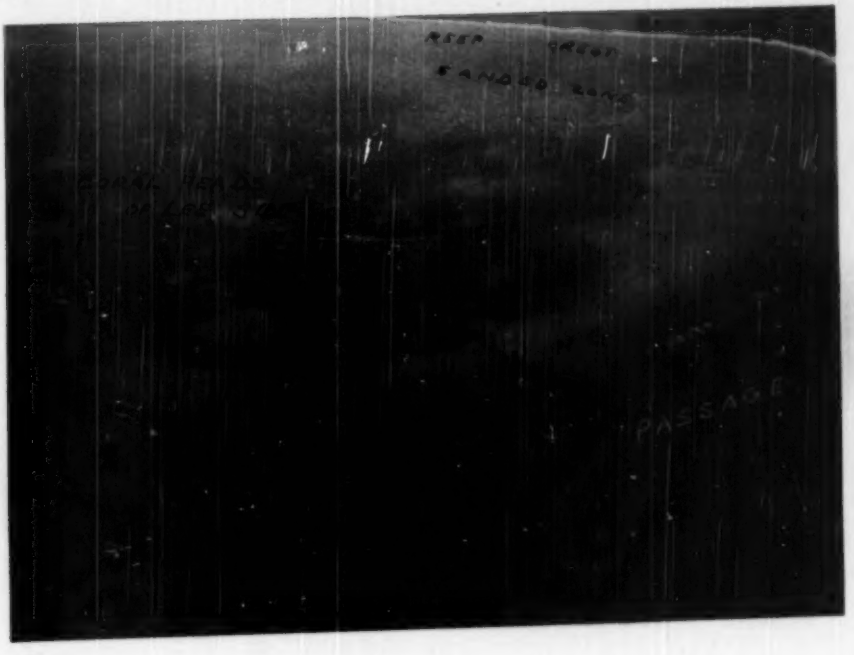
A



B

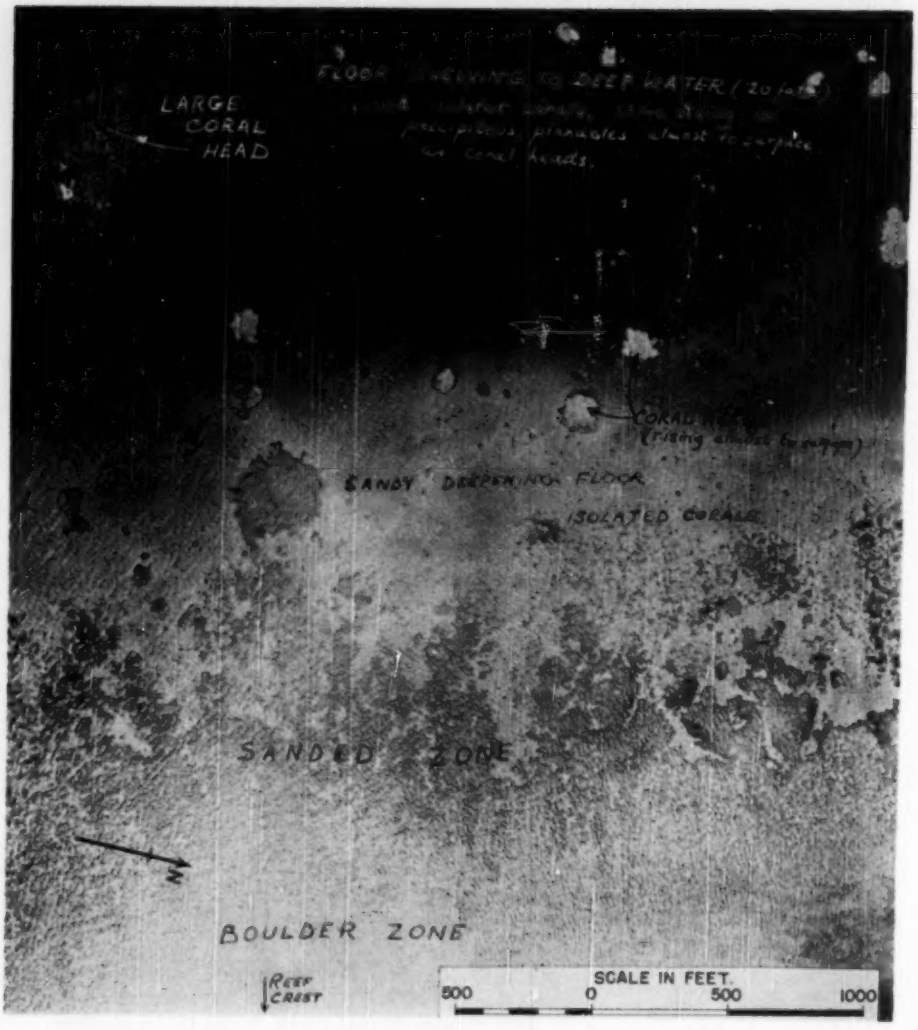
Outer Barrier Reef

A



B

Outer Barrier Reef



Outer Barrier Reef

southern coast to Albany (35° S., 118° E.) and beyond, there are no true coral reefs, but, instead, there is an inorganic erosional structure, commonly called a "sandstone reef."[3] In certain sectors there are four or five parallel reefs, spaced 3–5 miles apart. The outer ones are submerged, whereas the inner ones are in many cases awash at low tide. They consist of "coastal limestone," Pleistocene eolianite rocks, i.e., calcareous sandstones, former sand dunes, of the type described by Sayles (1931) from Bermuda, in the south, and red Tertiary sandstones and other rock types in the north.[4]

These "sandstone reefs" are commonly confused with true coral reefs because they consist in part of carbonates, which, like coral, are soluble in the sea water of the intertidal zone and which become eroded into smooth, nearly horizontal "reefs" at or slightly above low-water level.[5]

[3] These "sandstone reefs" are a peculiarity of the Western Australian coast line that are hardly to be found duplicated on the same scale anywhere in the world. The great sandstone reef of Brazil is the nearest analogue (Darwin, 1842; Branner, 1905). Those of Australia, however, are much longer and more diverse in character.

[4] These structures with patches and veneers of corals resemble, to some extent, the reefs of Bermuda (Nelson, 1837; Agassiz, 1895; Verrill, 1900, 1905; Sayles, 1931), those of the Bahamas (Nelson, 1853; Agassiz, 1894; Vaughan, 1914*b*), and also those of the eastern Mediterranean, such as at Alexandria (Crossland, 1907*b*).

[5] The question of the chemical action of intertidal sea water has been discussed by Macfadyen (1930), Kuenen (1933), Fairbridge (1948*a*, 1950*c*), and the physiography of these "sandstone reefs" is exemplified by the coastal limestone eolianites at Point Peron, Western Australia, by Fairbridge (1950*c*).

The reefs commonly support quite an active and varied growth of corals, calcareous algae, and the usual coral-reef invertebrate life and flora, thus giving strength to the mistaken idea that they are true coral reefs. Reef-building coral genera are found sporadically in the favorable environment of these "sandstone reefs" as far as 34° or even 35° S. latitude.

In many places the outer margins of the reefs are protected by a *Lithothamnion* ridge similar to normal coral reefs, and in certain localities the ridge is made up of a series of algal-rimmed water terraces of the type described and illustrated by Lister (1891) and Hoffmeister (1932) from Eua (Tonga); by Umbgrove and Cosijn (1931) from southern Java (see also Kuenen, 1933); and by Stearns (1941) from Guam and (1945) from Tongatabu. Examples along the Western Australia coast may be seen at Point Peron (Fairbridge, 1950*c*); at Cape Vlaming, Rottnest; between Dongara and Geraldton; and at the mouth of the Murchison River, where they form micro-atolls.

Between North-West Cape, Point Coates, Maud Landing, and Cape Farquhar (23°36′ S.), for a distance of 125 miles, there is an almost continuous ribbon-like "sandstone reef" interrupted by a few narrow passages, lying about 1–3 miles offshore. Inside, there is a "lagoon" with depths to 7 fathoms. It is a pseudo-barrier reef, 2,000–5,000 feet in width, exposed as much as 5 feet at low tide. It is fringed by vigorous coral growths, which almost block the lagoon in places. The reef foundations seem to be locally red Tertiary sandstone, elsewhere old reef limestone or eolianite.

Along the coast from Red Bluff to Cape Cuvier and Charles Point, reefs form a fringe up to 3 miles wide, but it is not known how

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PLATE 3

A, Oblique air photograph, looking southeast along outer-barrier reef (at 15°01′ S.), some 5 miles north of Lark Pass. The contrast between the surf-covered outer margin and the calm-water conditions of the lee side with overhanging type of coral growth is clear.

B, Curving wing, characteristic of the extremities of the outer-barrier reef morphology. Oblique air view shows southern end of the reef north of Carter Reef, with Cook's Passage in foreground (Captain Cook passed through here in 1770). Note deep water in the passage (15–20 fathoms), whereas the shallow (3–8 fathoms) water to the lee side favors growth of coral heads. (R.A.A.F. official photographs, not to be reproduced without permission.)

PLATE 4

Vertical air photograph, showing detail of the leeward (concave) side of an outer-barrier reef that shows typical gradation from the boulder zone and sanded zone of the reef, with its growing inner fringe of coral, to the zone of isolated coral heads, growing up as isolated pinnacles from the bottom as the sandy floor slopes gradually down to 15–20 fathoms. (R.A.A.F. official photograph, not to be reproduced without permission.)

much of it is coral. The shores and islands of Shark Bay, including Bernier, Dorre, and Dirk Hartog Islands, are fringed with "sandstone reefs" showing local coral patches. Discontinuous "sandstone reefs" are found all the way to Cape Leeuwin. Thus they are known at Gantheaume Bay (Murchison River mouth), Port Gregory, Geraldton, Cape Burney (Greenough River mouth), Leander Point (Port Denison, near Dongara), Jurien Bay, Wedge Island, Lancelin Island, mouth of Moore River, Waterman's Bay, Trigg Island, Cottesloe, Fremantle, Rottnest Island, Garden Island, Point Peron, Mandurah, Bunbury, Cape Naturaliste, Hamelin Island, and near Cape Leeuwin. Most of these reefs I have seen. The numbers of coral patches grow smaller and less varied as one goes south. The last big coral patch occurs on Rottnest at 32° S. (Teichert, 1950).

BARRIER REEFS

The only barrier reefs in Australia are those of the outermost belt of the Queensland reefs. Some small fringing reefs around islands within the main lagoon, e.g., Lizard Island, may perhaps be regarded as incipient barriers. The discontinuous reefs near the edge of the Sahul Shelf may also be genetically related to the barrier type, but morphologically they are only patch or platform reefs (Teichert and Fairbridge, 1948).

Within the Great Barrier Reef, extending about 1,200 miles from near the Gulf of Papua to near Sandy Cape (roughly 9°–24° S. latitude), two main types of outer-barrier reefs may be recognized, namely, "ribbon reefs" and "platform reefs."

Ribbon-reef zone.—This zone extends from near Murray Island in the north to Trinity Opening opposite Cairns (about 16° S.). Here are ribbon-shaped reefs, the "linear reefs" of Jukes (1847), arranged in a festoon along the outer edge of the continental shelf, rising apparently from just inside the 100-fathom line. The edge of the continental shelf here is extremely precipitous, sloping in 2 miles to 600 fathoms and in places to over 1,000 fathoms. The reefs are therefore exposed to the Pacific. On the western, or landward, side is the gently shelving sand-covered floor of the Great Barrier Reef lagoon, 20–30 fathoms in depth (pls. 2–4).

The ribbon reefs, 1,000–1,500 feet wide and 2–15 miles long, are separated from one another by narrow passages, generally half a mile wide. Seen from the air, they have the appearance of a continuous rampart. Close inspection shows that, at the end of each section, the reef edge curves back into the lagoon like the horn of a sand spit (fig. 2).

Yonge Reef, one of the outer barrier reefs and located east of Lizard Island, has been studied by the Great Barrier Reef Expedition (Stephenson, Tandy, and Spender, 1931). This reef consists of an outer, parapet-like fringe, covered with living coral, and furrowed by wedge-shaped channels or "chutes." Inside the margin there is a shallow zone, the so-called "outer moat" of Stephenson, occupied by irregular patches of living coral, negroheads, and boulders, heavily washed by surf. From here the reef rises gently to the reef crest, the "*Lithothamnion* rim," 2–5 feet above low-water springs. The reef material consists of the debris of broken corals, boulders, shells, and sand, cemented together and generally encrusted by the calcareous algae *Lithothamnion*, which can stand a certain amount of exposure during low tides in calm weather. Large negroheads are commonly perched on the top of this crest, and there is, in some places, a definite boulder zone. The wider parts of the reef, where the sediment debris and algae form a definite pattern, normal to the reef edge, has been called the "trickle zone" or "radial zone." Inside the rim

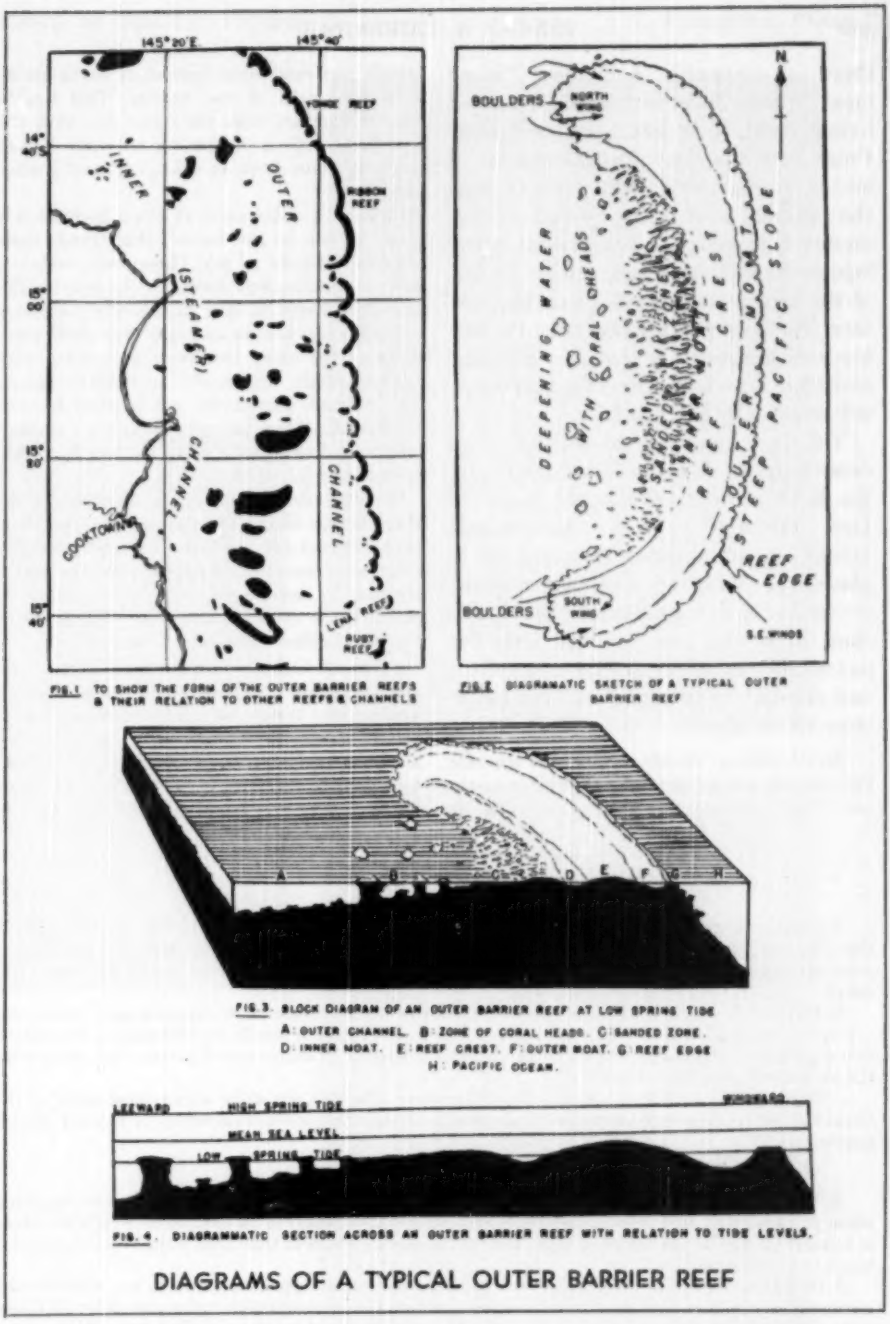

FIG. 2.—Diagrams of a typical outer-barrier reef, based on observations of Yonge Reef, Ribbon Reef, etc. (prepared by Dr. F. W. Whitehouse).

there is commonly a shallow "inner moat," partly filled with small clumps of living coral, boulders, and sediment. From here the floor deepens gradually and is covered with white coral sand of the "sanded zone," interrupted by numerous coral patches. Farther out in the lagoon the waters deepen, and only a few of the larger coral heads reach the surface. The deep passages between the ribbon reefs are generally free of coral heads, and there is no indication that these gaps will eventually be closed.

This is a typical cross section of an outer-barrier ribbon reef, as shown by the brief account of another, south of Lark Passage (15°8′ S.) by Agassiz (1898). Local variations occur, as in places the horns are long and narrow, protecting a rich growth of coral heads that, in places, fuse to form irregular patch reefs. In others, the horns are short and stumpy, or they may recurve to enclose small lagoons.

An interesting "double barrier" lies opposite the coastal section from Cape Melville to Lookout Point, between 14° S. and 14°25′ S. It trends west-northwest instead of north-northwest like most of the barrier. This double barrier diverges from the outer one with triangular junctions embracing an outer lagoon 10–17 fathoms deep, 40 miles long, and 5 miles across.

About 10 miles north of Cape Melville is a short section of the barrier that trends east-west (at latitude 14° S.). Three reefs, each 2–3 miles long, annular in form, are elongated in the east-west direction. The most westerly encloses a fairly deep lagoon and is a true shelf atoll, whereas the other two show a sandflat with shallow pools, apparently an infilled lagoon. The western extremities are marked by low sand banks. These lagoons, though not marked on the hydrographic charts, are clearly visible from the air (pl. 5, *A*).

On the ribbon reefs south of about 14° S., there are no islands or sand cays, but north of here there are a few unvegetated cays, generally situated at one extremity of the reef. The northernmost is Anchor Cay at 9°22′ S., north of which is the debouchment of the Fly River of Papua (Haddon, Sollas, and Cole, 1894).

Near 12° S. are several unusual features. The normal ribbon reef is breached by an opening leading into Wreck Bay, a circular expanse of water about 10 miles across, bounded by a ribbon reef, yet exceptionally deep (unbottomed by a 285-fathom line). About 15 miles north of here is a large atoll-like reef, 5 miles

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PLATE 5

A, Oblique air view, looking northeast over a section of the outer-barrier reef (14° S.), 10 miles north of Cape Melville, where the trend is locally east-west. Instead of ribbon reefs, there are three annular reefs, the most westerly a true shelf atoll, the other two with central areas apparently filled in with sand and coral debris.

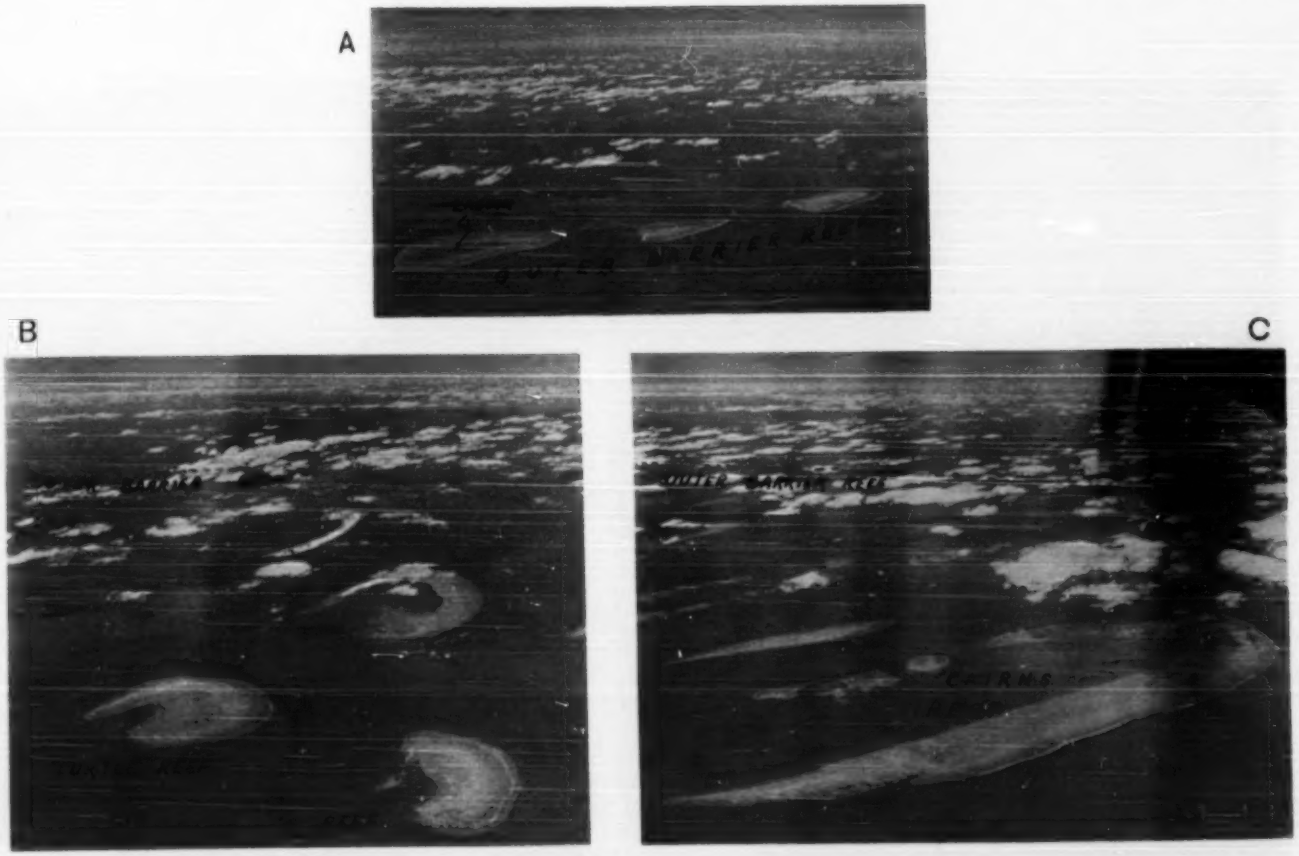
B, Oblique air view, looking east, showing characteristic horseshoe- or horsehoof-shaped inner reefs (15°25′ S.), including Turtle Reef, "E" Reef, etc. Note steep, smooth margin on southeastern (windward) side and shallow, irregular northwestern (leeward) side. All show an incipient atoll pattern, but are open on the lee side and partially infilled.

C, Oblique air view, looking east, showing Cairns Reef (15°40′ S.), one of the largest "near-atolls" of the Great Barrier Reef lagoon. It is 10 miles long and fails to close at the leeward extremity. (U.S.A.A.F. photographs.)

PLATE 6

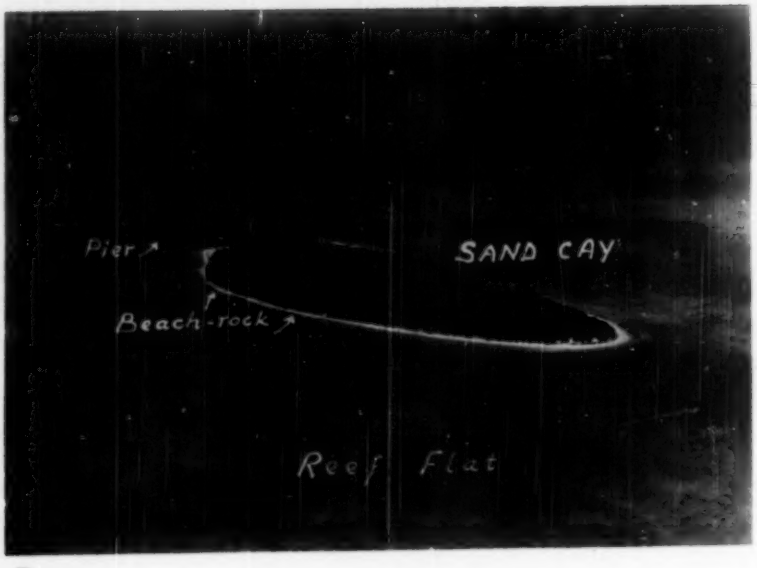
A, Vertical air photograph of Green Island (16°46′ S., 145°58′ E.), a typical vegetated sand cay, lying about 10 miles from Cairns. The cay lies at the west-northwest extremity of the reef. To the west of the fringe of living coral may be seen the usual sandy floor of the leeward sediment train from which rise isolated coral heads in 3–5 fathoms.

B, Oblique air view of Green Island, showing well the streamlined appearance of such cays, which consist purely of wave-accumulated sand but are partially protected from destruction during storms by the fringe of beach rock, which builds up rapidly beneath the surface layers of beach sand. (R.A.A.F. official photographs, not to be reproduced without permission.)



Outer Barrier Reef (A), Inner Reefs (B), and Cairns Reef (C)

A

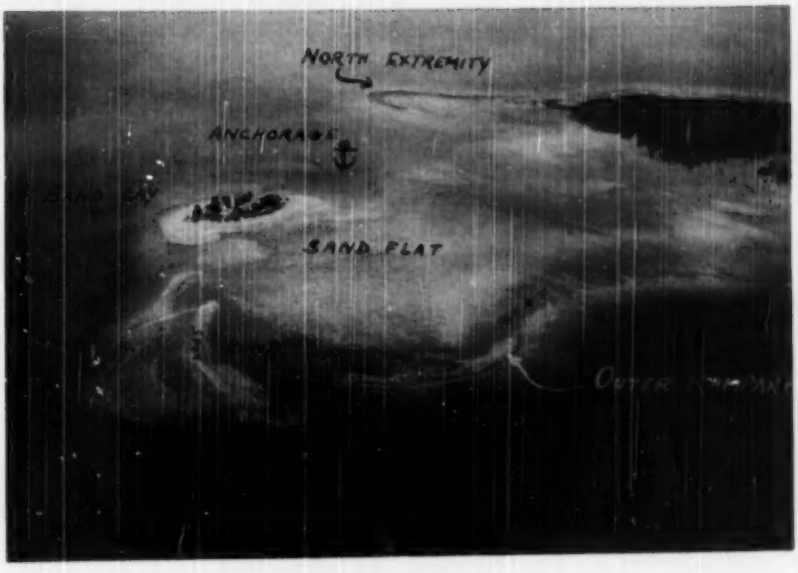


B

Green Island

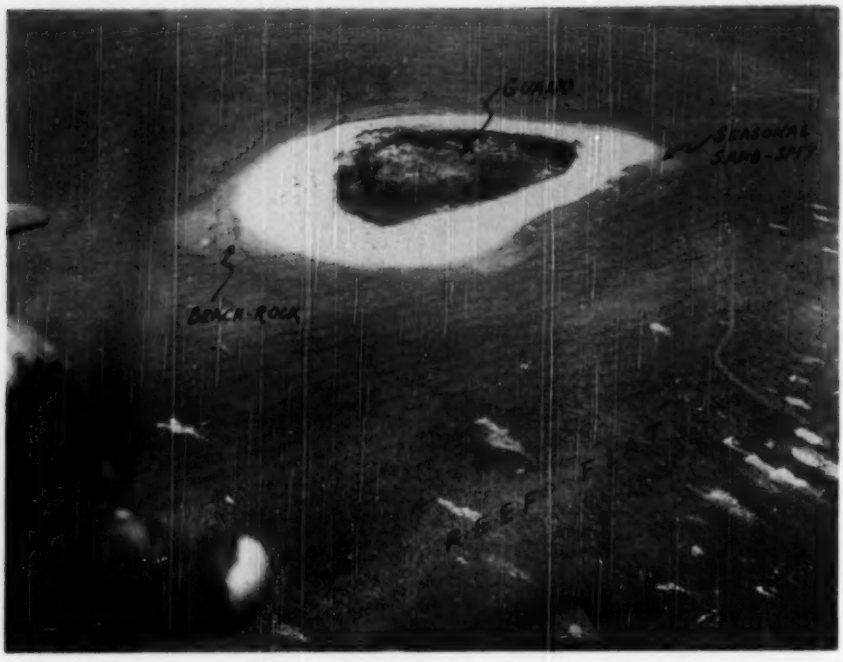
MAINLAND
SNAPPER ISLAND
SAND CAY
WEST EXTREMITY

A

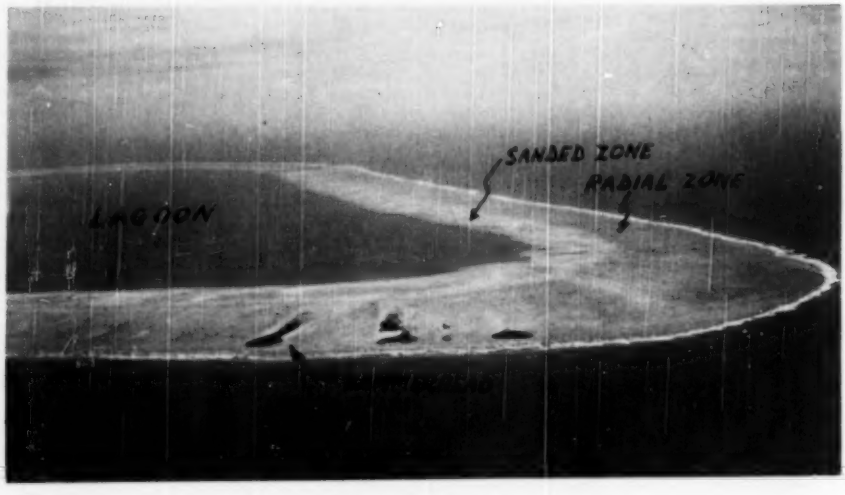

NORTH EXTREMITY
ANCHORAGE
SAND FLAT

B

Low Isles



A



B

Browse Island (A) and Seringapatam Atoll (B)

outside the outer barrier and separated from it by a channel over 100 fathoms deep. Five miles farther north there is an isolated reef platform with offshore slope of 45°–55°, surmounted by an island called Raine Island (11°35′ S.), the only one, I believe, in the outer-barrier region (Jukes, 1847).

Barrier platform reefs.—South of about 16° S. latitude, there is a distinct change in the character of the outer reefs. The ribbon reefs of the north are replaced by irregular-shaped patch or platform reefs. They are commonly set back several miles from the 100-fathom line and appear to be separated by deeper channels. The morphologic and ecologic zones recognized on the ribbon reefs are here less clear and, in any case, not well known.

Along the southern section of the outer barrier, islands or emerged coral platforms are found only in the extreme southeast. There we find Hixson Cay (22°30′ S.) and Bell Cay, with "insignificant sand patches," according to Saville-Kent (1893), on the south of the Swain Reefs, where the 100-fathom line lies more than 30 miles beyond the edge of the barrier. South of the broad Capricorn Channel are the scattered reef patches and islands of the Capricorn and Bunker groups. Morphologically, these are very similar to the island reefs of the inner platforms and are unlike the normal barrier types. They will be discussed under the heading "Reefs with Islands."

ATOLL REEFS

The writer is using the term "atoll" in an exclusively morphologic sense, that is, a ring-shaped ribbon reef enclosing a lagoon in the center. There seems to be no marked difference in appearance between an annular reef rising from an oceanic cone or platform and one rising from the continental shelf.[6]

[6] Davis (1928, p. 19) called the annular reefs which rise from relatively shallow platforms, "bank atolls." By "bank" he clearly meant such structures as continental or insular shelves, and not those features of the shelf such as sand banks, e.g., Dogger Bank in the North Sea. Molengraaff (1930, p. 56) considered these atolls as a variety of shore or fringing reefs because their foundations do not descend into deep water, and he called them "pseudo-atolls." This term, however, had already been used by Agassiz and others for platform reefs with shingle rims

PLATE 7

A, Oblique air view of Low Isles (16°23′ S., 145°34′ E.), looking north-northwest, showing the curious character of the "low wooded, island-reefs" which lie in the Inner ("Steamer") Channel, north of Cairns. The reef platform is fringed by a series of cemented shingle ramparts (in places marked by the mid-Recent erosion platforms of four thousand years ago), which protect the reef-flat and enable the establishment of mangrove, the roots of which are always awash.

B, Similar view of Low Isles, but looking northeast. Detail is shown of the small sand cay which characteristically forms from sand accumulated on the leeward end of such reef platforms. It lies almost midway between the two horns or wings of the horseshoe reef. Note the incurving spits of coral shingle on the two extremities of the reef. Shallow moats or ponds are protected within the shingle rampart system and support specialized types of shallow-water corals. (R.A.A.F. official photographs, not to be reproduced without permission.)

PLATE 8

A, Oblique air photograph of Browse Island (14°07′ S., 123°34′ E.) on the Sahul Shelf. It was probably first sighted by Dampier in 1699 and was exploited for guano in the late nineteenth century (traces of the workings may still be seen). This reef is an isolated, circular reef platform, mostly covered by waves of coral shingle (see Teichert and Fairbridge, 1948).

B, Part of Seringapatam Atoll (13°40′ S., 122°05′ E.), which lies near the western edge of the Sahul Shelf. Hammerhead coral-shingle spits may be seen on the radial zone in the foreground. An inner, sanded zone of the reef is distinguishable and probably represents the infilled portion of the lagoon. The reef drops to 200–300 fathoms on the exterior, so that some degree of Darwinian subsidence has probably taken place here. (R.A.A.F. official photographs, not to be reproduced without permission.)

The most perfect Australian atolls occur on the North-West Shelf and the Sahul Shelf. On the former, there are three major atolls. the Rowley Shoals (Mermaid, Clerke, and Imperieuse reefs), each being about 10 miles in length and surrounding a shallow lagoon (Davis, 1928, p. 231). Small sand cays are shown on them.

On the Sahul Shelf there are two: Scott Reef and Seringapatam Reef. Scott is a paired atoll because it consists of one annular reef in the north, some 10 miles in diameter, separated by a narrow but 100–400-fathom-deep channel from a horseshoe reef in the south, 18 miles across. Soundings show depths of 20–25 fathoms in the lagoon. A small sand cay surmounts a reef patch at its northwestern extremity. Seringapatam Atoll is an excellent, perfectly enclosed reef. It measures 6 × 4 miles, with a central lagoon about 20 fathoms deep (Teichert and Fairbridge, 1948, p. 241). In section, the reef is similar in character to the outer-barrier ribbon reefs of Queensland, except that it is about 3,000–4,000 feet wide. A well-marked zone of negroheads is found along the reef crest, and there are local accumulations of boulders and small shingle, forming "hammerhead" spits. The extra width of the reef occurs on the lagoonward side of the crest. Here, where waters flow over into the lagoon, there is a notable radial pattern produced by algae, boulders, and sediment—the "radial zone." Beyond this is a well-defined "inner zone" (up to 1,000 feet wide), carrying more sandy sediment, like the "sanded zone" of the ribbon reefs. Toward the lagoon this zone breaks up into coral patches which slope gently down into the fairly deep waters. A very small overflow channel crosses the southeastern corner of the reef (opposite the northwest-monsoon side). The height of the reef generally does not exceed 4 feet above low water, and, except for the shingle spits, there are no cays or islands.

These northwest atolls rise near the outer edge of the continental shelf, which here has the exceptional depth of 300 fathoms before it begins the steep continental slope to oceanic depths. Davis (1928) classified quite a number of the northwest platform reefs as atolls, for example, Adele, East Holothuria, Long Reef, and Lacepede. The last two enclose shallow lagoons ("pseudo-lagoons"), but the first two show not the least resemblance to atolls.

On the otherwise almost coral-free continental shelf of the southwestern part of Australia, we find an exceptionally large coral group, one of the most southerly (29° S.) in the Indian Ocean, namely, the Houtman Abrolhos group (fig. 3). The Abrolhos rise from the continental shelf, between 25 and 30 fathoms, and consist, from south to north, of four groups decreasing in size and extending over nearly 50 miles. These are the Pelsart, Easter, Wallaby, and Northern groups. They are complex reefs, irregular and discontinuous, but essentially annular and classified by the writer as shelf atolls (1948*a*). Some of the difficulty in classifying them arises from the fact that they have recognizable cores of erosion-truncated Pleistocene reef material, upon which are superimposed layers of younger material. Thus in the Pelsart group (Dakin, 1919; Teichert, 1947; Fairbridge, 1948*a*) Pleistocene coral limestone is marginally exposed on the islands but is partly overlain by and replaced in the interior by a lagoonal limestone, a white, calcareous mudstone, which is well exposed on islands rising from the lagoon floor. The early and mid-Recent deposits consist of shingle banks, blown sand, shell beds, and younger coral reefs.

Morphologically, the southwestern part of the Pelsart group, facing the swells of the Indian Ocean, is a ribbon reef, 12 miles long, with curving horns at each end like those of northern Queensland and having a similar cross section. However, it is uncertain what parts of the reef are "contemporary" and what are Pleistocene or early Recent. The southeastern part consists of the eroded platform and islands of Pleistocene coral rocks, fringed, both on the outside and on the lagoon side, by living corals. In the northern part, Pleistocene erosion has cut numerous deep channels, resulting in many small islands of emerged coral, fringed by vigorous contemporary reefs. The main lagoon is very irregular in depth, suggesting successive erosion platfoms developed during the Pleistocene.

The other groups in the Abrolhos differ morphologically from Pelsart but appear to have gone through the same history. In the Wallaby group, there are three small subsidiary atolls on the southern side: Morning, Noon, and Evening reefs, each being 2–3 miles

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and shallow reef-flat pools (see below). The term is not very satisfactory because it implies something that is not a true form. The present writer prefers the term "shelf atoll" because it means what it says: an atoll associated with a shelf area.

across. None of them shows any "elevated" coral limestone.

Atolls have not been recognized off northern Australia, but a number of "lagoon atolls" (Davis, 1928, p. 518) occur in the Great Barrier Reef lagoon of Queensland. These are intermediate between simple reef patches, horseshoe reefs, complete atolls, and ovoid platform reefs with the appearance of filled-in lagoons in the center. Saville-Kent (1893), Agassiz (1898), and Hedley (1906, etc.) have referred to the reef platforms with high shingle rims like those of the Capricorn and Bunker groups as "pseudo-atolls" or "pseud-atolls," enclosing "lagoonlets" or "pseudo-lagoons." This is not suitable nomenclature, as it is applicable only at half-tide!

Pickersgill Reef (15°52′ S.), the most perfect lagoon atoll, is an ovoid platform rising sharply

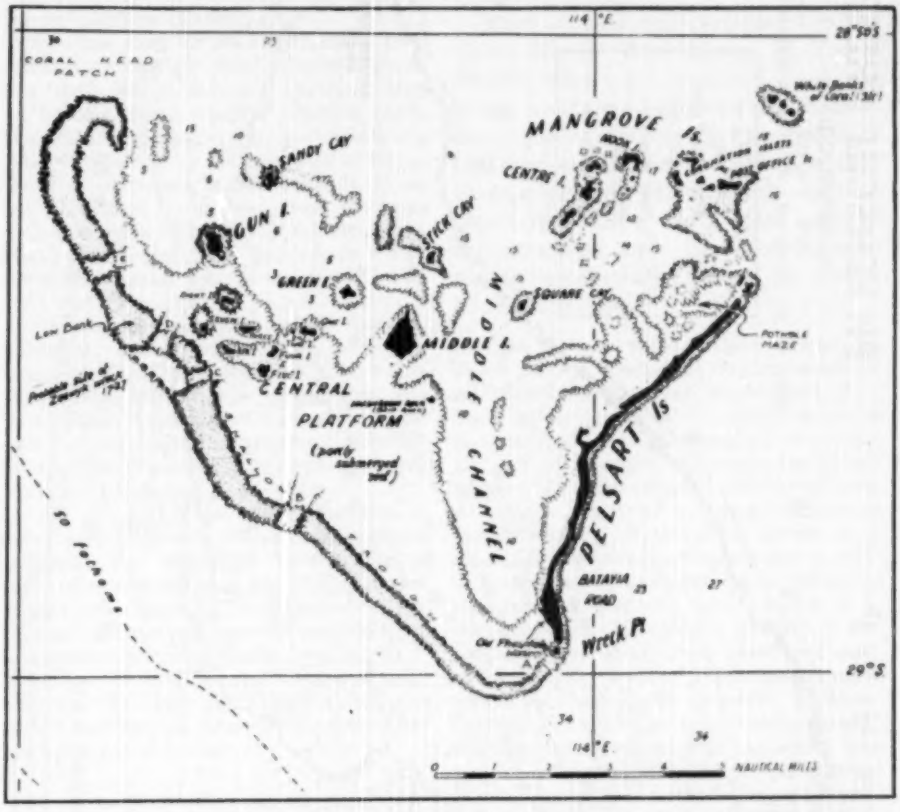


FIG. 3.—Pelsart group in the Houtman Abrolhos Islands, showing barrier type of southwest reef, bordering a platform of truncated Pleistocene reef and lagoon limestones. Islands of the latter are preserved along the southeastern margin and at scattered points in the lagoon, whereas the northern ("mangrove") islands are only mid-Recent emerged reefs.

spersed among the platform reefs of the mid-lagoon, and there appears to be a transition between simple reef patches, horseshoe reefs, complete atolls, and ovoid platform reefs with the appearance of filled-in lagoons in the center. Saville-Kent (1893), Agassiz (1898), and Hedley (1906, etc.) have referred to the reef platforms with high shingle rims like those of the Capricorn and Bunker groups as "pseudo-

from 18 to 24 fathoms (fig. 4). It is partly open on the west side. The "radial zone" is about 1,800 feet wide, sloping slightly to a sanded zone with shallow coral pools. This zone, in turn, generally drops off into a shallow lagoon, studded by numerous coral patches that rise from the floor and divide it in half. Although part of the lagoon is thus cut off, the southern part is open on the west. Depths of 3–4 fathoms

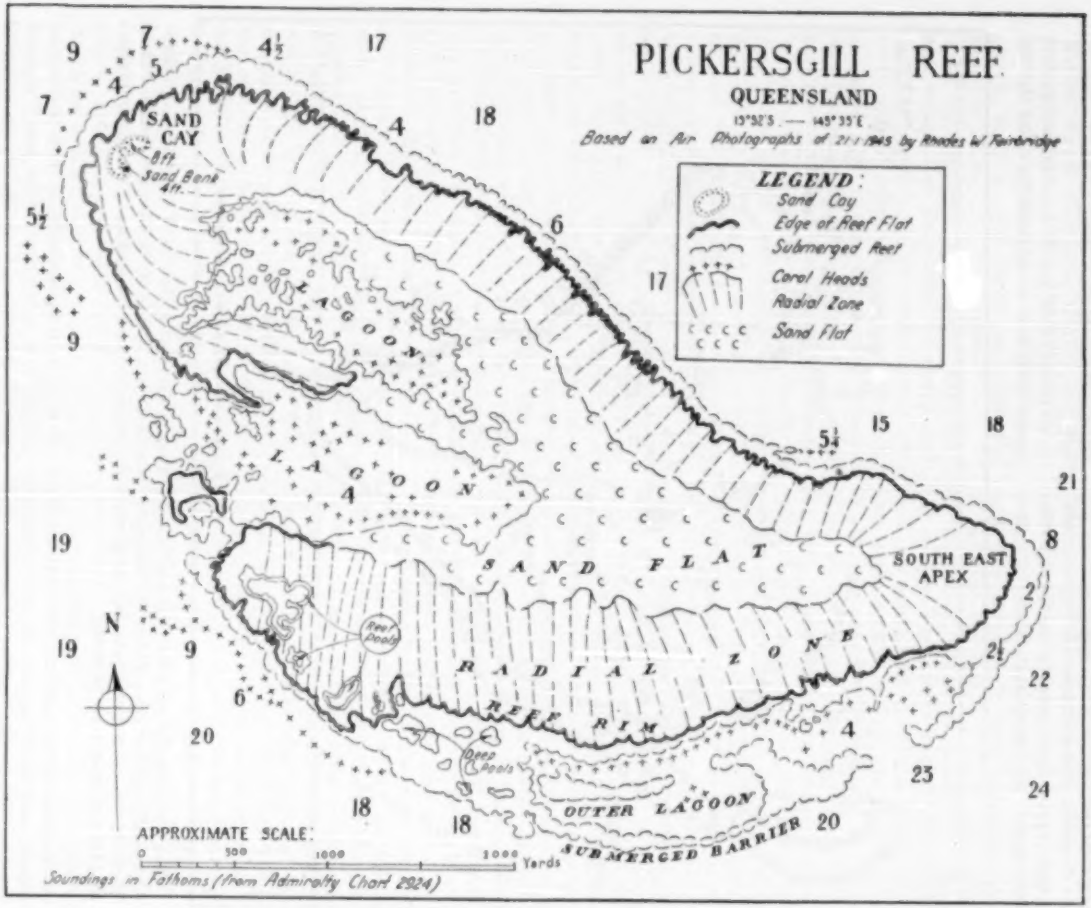


FIG. 4.—Pickersgill Reef, Queensland (based on air photographs)

were sounded in the lagoon (Steers, 1938). The southern side of the reef is fronted by an outer "microbarrier" lying about 1,500–2,000 feet from the reef edge. Patches of growing coral connect it with the main reef, but deep, rounded pools are left between. A sand cay, 300 feet long, rises 8 feet from the northwestern extremity of the reef platform (Steers, 1938).

A curious group are the Hardy, Hook, Line, and Bait reefs (19°45′ S.), 30 miles off Whitsunday Island (Edgell, 1928a). Rising from about 30 fathoms, there are two complete atolls, Hardy and Bait, and two near-atolls, Hook and Line. The latter are open to the northwest. Between the reefs, particularly Hardy, Hook, and Line, there is a channel about 500 yards wide and over 30 fathoms deep, with steep and locally overhanging sides. The lagoons within the atolls, studded with growing coral heads, are 5–10 fathoms deep.

There are numerous examples of crescentic, horseshoe-, and horshoof-shaped reef platforms, such as Cairns Reef (pl. 5, *C*), East Hope Island, Arlington Reef, Ellison Reef, and Trunk Reef. All are open to the north or northwest, and, because they are similar to the ovoid or irregular reef platforms, they will be referred to under that heading.

Only one reef of the Capricorn group, as noted above, encloses a sufficiently deep lagoon to justify the designation "atoll." This is Lady Musgrave, where depths of 4–5 fathoms occur in the lagoon. A very narrow passage through the outer reef is said to have been cut by Japanese fishermen (Steers, 1938, p. 56).

Sections of the Outer Barrier ribbon reefs, with their incurving horns, show a tendency toward atoll form, as noted above, in the sector north of Cape Melville.

PATCH OR PLATFORM REEFS

These are by far the most numerous of the reefs around Australia. This category includes all the irregular islandless reefs which generally rise from the shallower parts of the continental shelf. They are quite distinct from the barrier types, atolls, or fringing reefs. The term "patch reef" is applied to the smaller ones, and "platform reef" to the larger. The adjective "platform" means only that the reef is a broad, flat, tabular affair; it does not imply growth from an antecedent platform. It has no genetic significance.[7] This section discusses only patch or platform reefs *without* islands. Those with islands present special problems and will be discussed separately.

Most of these reefs are found in the Great Barrier Reef lagoon, in Torres Straits; others are scattered along the outer, northern edge of the Sahul Shelf; numerous others lie off the shores of northwestern Kimberley, in the Dampier Archipelago, and intermittently down to North-West Cape, where small patches are associated with the Western Australian "sandstone reefs" and the Abrolhos atoll groups.

These reefs are without islands and largely awash except at low tide. A number have been described from the Great Barrier Reef. Batt Reef (16°27′ S., 145°48′ E.), which we have reconnoitered and photographed from the air (pl. 1, *B*) and which Stephenson, Tandy, and Spender (1931) have mapped, may be taken as an example. This reef is typical of many in the central part of the Great Barrier Reef lagoon. It rises sharply from the lagoon floor in 15–25 fathoms, being steepest in the southeast and most gentle in the northwest. It is oriented northwest-southeast, parallel to the southeast trade wind, which blows here most of the year. The reef is 10 miles long and 4 miles across, the outer margin describing a smooth parabolic curve with the apex in the southeast. Only the northwestern edge is ragged and broken up by coral heads, apparently growing from a sediment cone on the lee side of the reef. The top of the reef is a flat platform without any lagoon. The reef crest rises to about 4 feet, and the central reef-flat stands at about

[7] Jukes, Dana, Stephenson, Spender, Steers, and others refer to these as "inner reefs," meaningless when there is no barrier. Niermayer (1911) named them *plaatrif*, which I translate as "platform reef." Molengraaff (1930) translated it as "shoal reef," but Teichert (1947) gave "shelf reef" or "hummock reef." Davis calls them "bank reefs." *Plaatrif* considerably antedates Tayama's "table reef" (1935), recently adopted by Ladd (1950, Am. Assoc. Petroleum Geologists Bull. 34, p. 204). (See n. 6, p. 341, for other reef terminology.)

1 foot above datum. Because the spring tidal range is about 9 feet, the reef is generally awash. Sand cays, shingle banks, or "raised" reef remnants are lacking. The reef surface is simple. From the reef crest, just within the breaker line, there is a broad, radially striated "trickle zone" of sediment debris and algae, which ranges from 4,000 feet in width at the southeastern apex to nothing at the northern end. It averages 2,000 feet along the sides. Within the trickle zone is a well-defined central sand-flat, consisting of a bare expanse of coral sand and small boulders, interrupted here and there by shallow pools fringed or carpeted with growing coral. Growing corals are also found all around the periphery of the reef and rising as irregular "heads" at the shallow northwest end.

Another large reef platform is Cairns Reef (15°20′ S.) (Hedley and Taylor, 1908). It forms a great horseshoe platform, 10 miles long, and open to the northwest (pl. 5, *C*). The outer edge is fringed with living coral; the 300-foot-wide crest is characterized by Alcyonaria. Behind the crest is a shallow sandy depression with small coral colonies, which grades into a broad sand-flat, with shallow, sandy pools containing isolated corals, Mollusca, *Halimeda*, and *Zostera*. Scattered negro-heads rising to 8 feet above datum occur along the windward rim. The bulk of the reef platform, 8,000 feet across, stands at only 1–2 feet above low-water springs. At the inner margin there is a gentle slope down to 5–10 fathoms, dotted with numerous coral heads.

One of the smaller platforms is Bee Reef, which lies only 1 mile off the northwestern tip of Cairns Reef (Hedley and Taylor, 1908). It is oriented northwest-southeast, an elongated oval, 1 mile in length. On the windward side is found a coral-shingle beach ridge, or rampart, 12 feet high. Mean spring tidal range here is 9½ feet. Behind the shingle bank is a long tongue, 100 feet long, in the form of a hammer-head spit. Shallow pools with living corals occur in the bordering 600 feet, followed by a sand-flat, 2,500 feet across, with fragments and slabs of dead coral. A very small, unvegetated sand cay marks the leeward end.

Numerous platform reefs occur in the Torres Straits, almost blocking the channel between Australia and New Guinea (Haddon, Sollas, and Cole, 1894). Air photographs show that they are similar to the large platform reef of the Great Barrier Reef lagoon but differ in shape and orientation. They are boat- or canoe-shaped, long, narrow, and pointed at each end, with an east-west orientation apparently related to the currents passing through Torres Straits rather than to wind directions.

A similar group of reefs is found in Clarence Straits, Northern Territory, where the orientation and shape of the reef platforms also seem to be controlled by the current directions (fig. 10).

The northern part of the Sahul Shelf is essentially free from coral reefs, except along its edge, which is characterized by a "much-interrupted barrier." The reefs—Cartier, Ashmore, Troubadour, Flinders, and Evans—all appear to be similar. All rise from inside the 100-fathom line; they cannot be distinguished morphologically from the platform type. Cartier alone has been studied with the aid of air photographs (Teichert and Fairbridge, 1948). It has a small sand cay that is unvegetated and may be entirely submerged in heavy weather. The reef platform, a smooth ellipse, owes its form to waves and currents that undergo a complete reversal each monsoon season. The reef, about 1½ by 3 miles, has a radial zone of sedimentary debris and algae extending almost to the center, where there is a restricted sand-flat with shallow, coral-fringed pools.

In this same region are several large reefs, such as the Sahul Banks and Lynedoch Bank, that do not break the surface; they are covered by 10–15 fathoms of water. It is not certain whether or not they contain living coral. For the most part, they appear to have been "drowned" by subsidence.

The numerous platform reefs on the interior of the Sahul Shelf off the rocky northwest Kimberly coast, such as the East and West Holothuria reefs and Long Reef, rise from submerged banks or terraces about 15 fathoms in depth. Small, submerged coral patches rise from these banks. Farther southwest, in the outer parts of the Buccaneer Archipelago, there are islandless platform reefs and patches, such as Churchill Reef, rising from banks which are 15–25 fathoms above the general level of the shelf, here covered by 35–45 fathoms of water. Many of these submerged banks are covered by dense forests of Bryozoa and Alcyonaria rather than coral. They grow down to 30–60 fathoms (Basset-Smith, 1899). This area is also one of the world's great pearl fisheries.

The North-West Shelf contains only numerous small patches close to rocky parts of the coast line, between the Dampier Archipelago,

Monte Bello Islands, Barrow Island, and North-West Cape.

From North-West Cape southward, only isolated coral patches occur near the shore. In the lagoon of the Maud-Cloates sandstone reef and in the Abrolhos region (Dakin, 1919; Teichert, 1947; Fairbridge, 1948*a*), there are, however, luxurious growths within the protecting barriers.

REEFS WITH ISLANDS

Many Australian coral reefs possess islands, the material of which consists of calcareous fragmental coral, molluskan, algal, or foraminiferal sand or coral shingle. In contrast, islands with mainland bedrock are designated as "continental" or "high" islands by Steers (1929) or as "hilly timbered islands" by Hedley and Taylor (1908). Those surrounded by fringing reefs are grouped separately.

True "coral" islands have been classified according to types. Spender (1930) distinguishes five classes: (1) islandless reefs with scattered debris only; (2) reefs with a sand cay but no shingle ramparts (peripheral beach ridges of coral shingle); (3) reefs with sand cay and ramparts but with no vegetation on the latter; (4) reefs with sand cay and ramparts, with vegetation on the latter; and (5) reefs with all the features of class 4, together with a mangrove swamp on the reef-flat.

This system is somewhat artificial, as no definite example of class 4 is known, and there are only two small examples of class 3. Furthermore, it does not provide a category for islands exposing cores or platforms of older coral-reef material, such as Raine Island in the Outer Barrier of Queensland, some of the Abrolhos islands of Western Australia, and certain larger islands of the Sahul Shelf. Spender (1930) named his class 5 "island-reefs," excluding reefs with large vegetated cays, such as Green Island and Heron Island, an arbitrary restriction. Steers (1929, 1937) adopted Captain Cook's term "low wooded island" for Spender's class 5. Again, this is not entirely accurate because large vegetated cays are also "low," "wooded," and "islands." Steers (1937) further subdivided the latter into those mainly with sand and beach rock and those prevalently with shingle.

We have adopted here a revised classification of coral island types found on the Australian shelf. This is as follows:

Type 1—sand cay, unvegetated.—Commonly unstable and migrating seasonally. Found in all areas on all except the fringing reefs.

Type 2—sand cay, vegetated.—Moderately stabilized, generally with beach rock. Widely distributed but generally missing from the outer, more exposed reefs.

Type 3—shingle cay, with or without vegetation.—Moderately stabilized, widely distributed, generally found on smaller, more exposed reefs.

Type 4—sand cay with single ramparts (beach ridges).—Unvegetated to completely vegetated islands, including mangrove swamp. Stabilized. This category overlaps type 5, as small "pavements" or "promenades" (i.e., benches) of older emerged reef-core material are commonly associated with the "rampart" system (Steers, 1937). Restricted to smaller reef platforms in protected parts of the northern Great Barrier Reef lagoon.

Type 5—island with exposed platform of older, emerged coral-reef material.—With or without a fringe of recent sand or shingle beach ridges or ramparts. Platform vegetated and stable (fig. 5).

The following described examples may serve as illustrations of the redefined categories:

Type 1.—Numerous examples are found in the Great Barrier Reef lagoon, such as Sud-

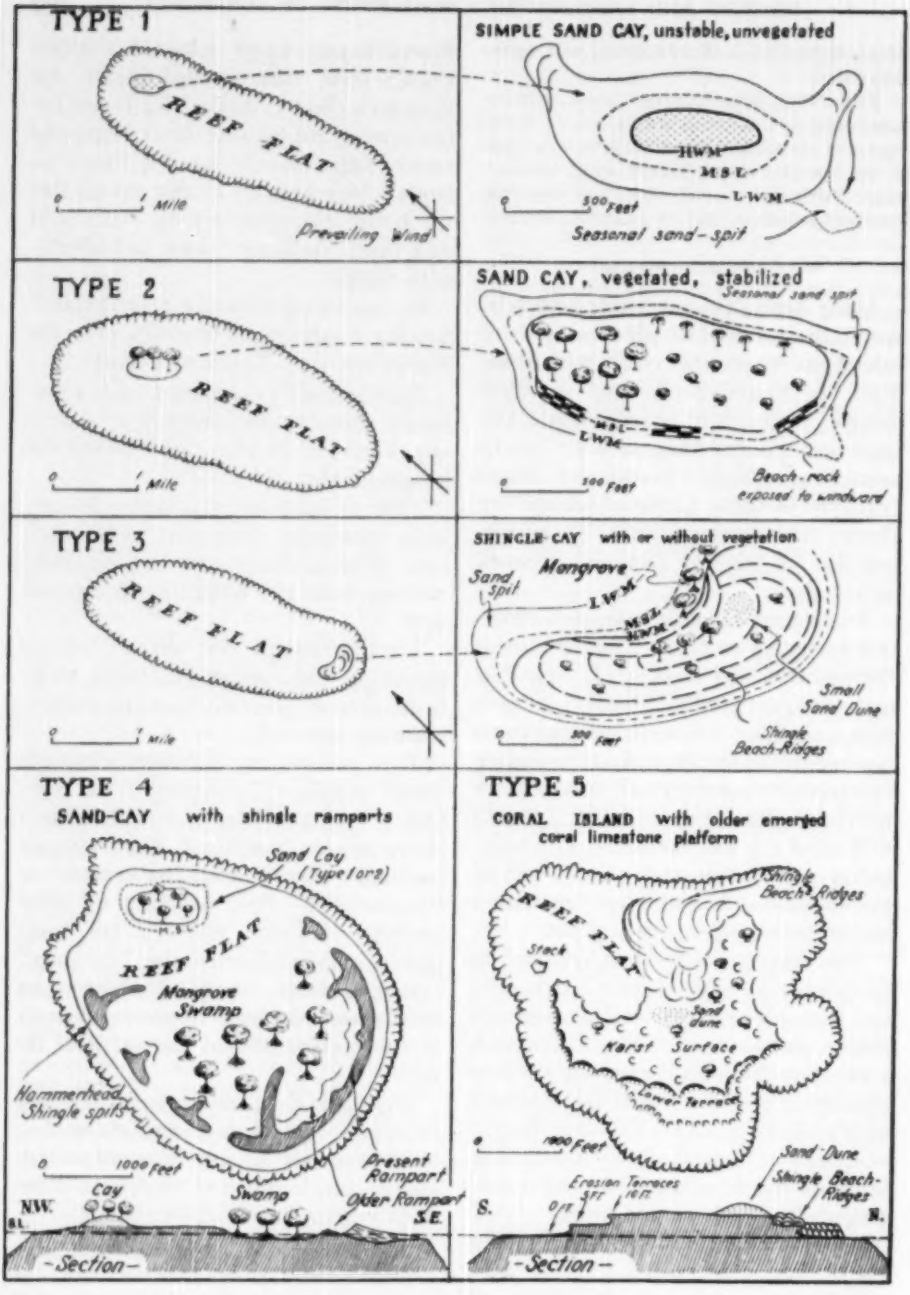


FIG. 5.—Coral-island types from the continental-shelf areas of Australia

bury, Undine, Mackay, Pickersgill, Chapman, and at least eighteen cays on the outer barrier north of 16°13′ S. (Steers, 1937, 1938). Other examples occur on the North-West and Sahul shelves, e.g., Cartier (Teichert and Fairbridge, 1948).

Type 2.—Some of the best-known coral islands belong to this class, such as Heron Island, North-West Island, Tryon Island, Erskine Island, Mast Head Island, and others in the Capricorn group (Steers, 1937, 1938). Other characteristic examples are found on Arlington Reef, Michaelmas Reef, Green Island, Pelican Island, Fife Island (Steers, 1938), Combs Cay, Stapleton Cay (Steers, 1929), situated at intervals farther north along the Great Barrier Reef lagoon. Adele Island, on the Sahul Shelf, probably belongs to this class (Teichert and Fairbridge, 1948).

Type 3.—Lady Musgrave Island, Lady Elliot Island, One Tree Island, and others in the Bunker and Capricorn groups are characteristic examples (Steers, 1938). Representatives in Western Australia are found in the Abrolhos group (Fairbridge, 1948*a*) and Browse Island on the Sahul Shelf (Teichert and Fairbridge, 1948).

Type 4.—This category comprises complex varieties of coral islands, known only in the Inner, "Steamer," Channel of the northern sector of the Great Barrier Reef lagoon, the "island-reefs" of Spender, and the "low wooded islands" of Steers. Examples of Spender's class 3, the simplest variety, without vegetation on the ramparts, are Bee Reef (Hedley and Taylor, 1908) and one of the Turtle group (Steers, 1937, p. 137). The transition to the next variety, Spender's class 4, where some vegetation, generally *Avicennia*, colonizes the ramparts, is seen in East Hope Reef (Steers, 1937). Finally, there are the numerous examples, class 5, where cay, ramparts, and reef-flats are vegetated and the reef-flats are converted into a mangrove swamp by *Rhizophora*, such as Low Isles (Stephenson, Tandy, and Spender, 1931; Fairbridge and Teichert, 1947, 1948), Three Isles (Stephenson, Tandy, and Spender, 1931), Dhu Reef, Ingram and Beanley Islands, Pipon Island, Night Island, Houghton Island, Croquet Island, Bewick Island, Low Wooded Island, King Island, Enn Island, Newton Island, Watson Island, Turtle Island, Kew Island, Hope Island, Two Isles, and others (Steers, 1938).

Type 5.—Coral islands consisting principally of Pleistocene to early Recent reef material are rare in the Australian region because of the lack of tectonic uplift and the rapidity of chemical solution, both subaerial and intertidal. An example is Raine Island (Jukes, 1847) in the Queensland barrier reef complex. This is principally an emerged platform. Steers (1929) placed it among the "continental" islands, but there is no evidence of a continental core. The near-shore reef islands of the North-West Shelf appear to be of this category, but detailed surveys are lacking. Examples in the southwest are the Houtman Abrolhos groups, with complex cores of Pleistocene reef and dune materials and younger covers of mid-Recent coral reefs, shingle beach ridges, and sand dunes (Teichert, 1947; Fairbridge, 1948*a*).

SUB-RECENT AND PLEISTOCENE REEFS

Owing to lack of major Quaternary tectonic uplift around the reef-fringed coasts of Australia, there are relatively few older elevated reefs. However, eustatic rises of sea level permitted the growth of coral reefs to heights well above the present low-water level.

Taking the "standard" eustatic levels recognized in the Mediterranean and on the eastern seaboard of the United States as our guide, we should find emerged reefs at 25, 100, 200, and more feet above sea level. Yet we have not found a "fossil" reef above the 25-foot late Pleistocene level. Even these are extremely rare, the bulk of the emerged reefs being limited to the mid-Recent 10-foot eustatic level. Because of the deeply weathered condition of these reefs, Pleistocene reefs originally present could have been completely destroyed by the rapid subaerial erosion that characterizes subtropical climates.

An alternative hypothesis is that during the Pleistocene the cold conditions, which had set in during the glacial stages, inhibited the growth of reef-builders in the marginal areas (on the Queensland coast from 10° to 25° S., for example), *even during the excessively warm*

interglacial periods, when melting of the icecaps caused the sea level to rise 50–100 feet or more above the present mean. The writer finds neither hypothesis completely satisfactory.

Almost the entire coast line of Australia shows traces of the mid-Recent high sea level, which seems to correspond with the "Atlantic" stage of western Europe, the climatic optimum of 3,000–4,000 years ago. Where the record is clearest, as in the Western Australian Pleistocene eolianites and "fossil" coral reefs, Teichert (1947, 1950) and the writer (1948*a*, 1950*c*) find evidence for two intermediate stillstands, the last being perhaps only five hundred years ago. These Recent eustatic levels occur at 10–11, 5–6, and 2–3 feet above low-water springs. Similar levels are found in the East Indies (Kuenen, 1933; Umbgrove, 1947) and elsewhere in coral seas.

At each stillstand, pre-existing reef limestones should be cut into by a marine bench, and new fringing reefs should grow up to the height of the corresponding low-water spring level. Normally, corals do not grow more than 1–2 feet above low-water springs, and erosional benches are cut in limestones by intertidal solution down to within 1–2 feet of low-water springs, thus providing a very satisfactory datum for all measurements.

Limestone platforms are therefore of coral grown *in situ* in places covered by growing corals or of older reefs or limestones truncated by erosion. On stable shores these platforms can be correlated with one another by means of elevation alone if they are spaced close enough to exclude doubt as to their identity. They are commonly associated with raised beaches which contain characteristic faunas.

These methods have been applied on the Western Australian coast, in the Abrolhos (Teichert, 1947; Fairbridge, 1948*a*), at Rottnest (Teichert, 1950), and at Point Peron (Fairbridge, 1950*c*), where studies were favored by the very low tidal range (generally under 4 feet and often less than 12 inches). Benches 10 feet above datum are here exposed only to spray or rain-water solution, whereas the 5-foot and 2-foot levels are commonly submerged and less well preserved than the older platform. In areas of great tidal range, *all* the benches undergo marine erosion.

Induration in the intertidal zone during their development greatly retards the later subaerial erosion of these benches. Solution and reprecipitation of $CaCO_3$, under alternate wetting and drying in the supersaturated sea water, produces an indurated crust on the calcareous material which is welded to a rock of almost granite-like hardness. This crust acts like a shield of armor plate on each bench, protecting it to a remarkable degree from subsequent erosion.

There are few precise surveys of these sub-Recent "elevated" reefs. Although there are diverse interpretations scattered through the literature, all seem to indicate one thing, namely, a slight recent emergence. There is some difficulty in evaluating the published data which differ in accuracy and selection of reference planes, such as "about the height of high neap water" or "corresponding to the lowest limit of living rock oysters." The writer has attempted to reduce the elevations to a common datum—the low-water springs level—in order to determine whether or not the levels conform to a recognizable eustatic pattern.

Western Australia.—The first of these studies of "elevated" reefs was undertaken in the southwestern part of Western Australia, where there is clearest evidence. Raised beaches and terraces have long been known there (Somer-

ville, 1919; Aurousseau and Budge, 1920; Jutson, 1934; Teichert, 1947, 1950; Fairbridge, 1948*a*, 1950*c*). Both older and contemporary benches in this area are "sandstone reefs," the older levels containing large colonial-type corals which are not found at present living within several hundred miles of here. Their "tropical" or "Indian" character was first recognized by Tenison-Woods (1878). The warm-water character of the associated molluskan fauna was pointed out by Reath (1925).

In the Abrolhos Islands, the oldest recognized reef limestones associated with white chalky mudstones ("lagoon limestones") are believed to be late Monastirian (Riss-Würm interglacial). On Middle Island these reefs rise to over 18 feet above datum (Fairbridge, 1948*a*) and were originated in a sea at least 25 feet above the present level. On West Wallaby Island they are overlain by 30 feet of dune limestone (Teichert, 1947), apparently corresponding to the younger eolianites of the "coastal limestone" of the mainland and probably representing the eolian interval during the lowered sea levels of the last (Würm) glaciations. Several fossil soil horizons indicate interruptions in the sand-dune accumulation. In early Recent times the sea level rose again, marked by coral reefs rising to 10 feet above datum. These are in part eroded down to lower platforms and in part veneered by younger coral reefs of the 5- and 2-foot sea levels. Superimposed on the reef platforms are coral-shingle beach ridges left by the retreating sea.

The most southerly elevated reef so far observed is at Salmon Bay, Rottnest Island (Teichert, 1950), and that of the mainland coast occurs at Point Leander, near Dongara, in latitude 29°18′ S. (Hartmeyer, 1907). Another one was recently located at Cape Burney, 6 miles south of Geraldton. Cross sections show the zoning relations between the reef-building corals (*Acropora*, *Platgyra*, *Favites*, etc.), and the encrusting *Lithothamnion* layers. In parts of the reef the lithothamnioids make up 80 per cent of the volume, but, in general, they form the upper part of the reef, corresponding to the former shallowest-water zone. These reefs are older than those of the early "10-foot" bench and, like the older Abrolhos reefs, probably Monastirian in age.

Similar "raised" benches and reefs occur along the coast as far as North-West Cape (Condit, 1935) and have been reported between Onslow and Cape Locker (Raggatt, 1936), between Port Hedland and Broome (Clapp, 1926), at Roebuck Bay (Broome), by Hardman (1884) and Cadell (1899), and in the northwest Kimberley and its offshore continental islands (Basset-Smith, 1899; Basedow, 1918).

Northern Territory.—A "raised" coral limestone bench, 10 feet above low-tide level, commonly associated with raised beach rock and shell beds at its inner margin, extends discontinuously over hundreds of miles along the northwestern coast line (Brown, 1906; Basedow, 1916), recurring on Melville Island and along the northern and eastern coasts of Arnhem Land (Brown, 1908).

Gulf of Carpentaria.—On Sweer's and Bentinck Islands, Jackson (1902) recorded a cap of Pleistocene eolian calcareous sandstone; and raised corals, probably of the 5- and 10-foot levels here, have been described by Jensen (1914) and Paradice (1924) from the Sir Edward Pelew Group.

Torres Straits.—There are few records of emerged coral platforms or associated beach deposits from this coral-studded region. The numerous studies in this region (Jukes, 1847; MacGillivray, 1852; Saville-Kent, 1893; Haddon, Sollas, and Cole, 1894; Mayer, 1918, and others) were centered either on bedrock or on the living reefs.

MacGillivray (1852, p. 32) referred to a raised coral conglomerate on the northern side of Nagir Island and on most of the other islands, as evidence of uplift. Haddon, Sollas, and Cole (1894) concluded that there were no signs of large-scale emergence or submergence in the Torres Straits, although they described emerged beach rock and recent coral conglomerate, which cannot be formed except in the intertidal zone.

Mayer (1918) observed: "In common with other islands of the Torres Straits region, whether volcanic, calcareous or continental in character, the Murray Islands exhibit a recently emerged shore platform about 3 feet above the present high-tide level. . . ." With a 7–8-foot tidal range, this means a 10–11-foot platform. An emerged coral reef at the same level was observed by Mayer on Thursday Island. There are analogous alluvial flats and raised beaches on many of the islands.

On Erub (Darnley) Island, Jardine (1928*d*) reported elevated terraces at about 12 feet above high-water level. Referring to Aurid, C. M. Yonge (1930, p. 184) observed: "This island, like many others in the Torres Straits,

shows unmistakable signs of comparatively recent elevation, the line of the old beach being easily distinguishable about 6 feet above the present one."

Haddon (1935) substantiated the evidence of emergence but concluded that it was not eustatic. Instead, he thought it was positive elevation still in progress. He said that the natives are certain that an elevation is now slowly taking place. Reefs formerly 12 fathoms deep are now estimated at only 8–9 fathoms, sand spits were building up, and wrecked vessels, formerly submerged, now appear above low-water level. Although Torres Straits appears to have been silting up for a number of years, this is no proof of elevation. In the absence of positive evidence of tidal gauges, emerging docks, and wharves, we should accept Haddon's interpretation with reserve.

Great Barrier Reef.—Raine Island (11°35′ S.), an isolated, emerged coral island on the Outer Barrier, consists of a platform of calcareous sandstone rising to 25 feet above low-water mark (Jukes, 1847). Containing fossil turtle eggs, it is, in part at least, an emerged beach rock. With a tidal range of 10–12 feet and an extra 2–3 feet to include mean swash, an emergence of about 10–11 feet is indicated. Rattray (1869) also mentioned a raised coral breccia here.

A definitely emerged coral reef occurs some 12 miles southwest of Raine Island (Jukes, 1847). It is 2–3 miles long and ¼ mile wide, with *Porites* colonies in the position of growth 10–12 feet above the present reef growth level.

There is ample evidence for a slight Recent emergence farther south along the Queensland coast and on many of the island and fringing reefs. As long ago as 1827, P. P. King referred to the "raised" beaches and other signs of emergence. Jukes (1847) noted the occurrence of banks of pumice all along the coasts of Queensland and New South Wales at a height of about 10 feet above high-water mark. He concluded that the latest movement of the Great Barrier had not been one of submergence.

In 1869, Rattray commented on the slight Recent emergence of the entire eastern seaboard. Guppy's map (1888) depicted a continuous raised reef ridge all the way from Cape York to Sandy Cape, elevated at least 10 feet. It was based on earlier reports and is certainly an assumption, as raised reefs along the mainland shore are discontinuous. Jack and Etheridge (1892) presented evidence of emergence along the mainland shore, principally, however, of emerged alluvial flats, raised beaches, and rock benches, although examples of emerged coral reefs were also cited. Stutchbury (1854) noted raised coral reefs in position of growth as far south as Peel Island in Moreton Bay.

Additional evidence of raised coral, beach rocks, and alluvial flats on the reefs, continental islands, and mainland shores has been presented by Stokes (1846), Agassiz (1898), Andrews (1902), Hedley (1925*b*), Jardine (1925, 1928*a*, *b*), Marshall, Richards, and Walkom (1925), Richards and Hedley (1925), Stanley (1928*a*), and Steers (1929, 1930, 1931, 1937) and has recently been summarized by Richards (1939*a*) and Browne (1945). Several of these workers recognized at least two Recent standstills.

Few authors agree on the height of present-day bench erosion. Steers (1937, p. 21) puts it "at or a few feet below mean sea-level" but was unable to prove present-day benching in Australia, owing to the comparatively recent date of the last movement. The writer agrees with Steers in the case of the harder rocks but feels that the evidence is clearer on the coral platforms, which are soluble in the sea water of the intertidal zone. Most "hard rocks" are slightly soluble in the carbonic acid–rich water, so that, given time, a present-day bench may be formed at low-tide level. This would be distinctly lower than Steers's position; but, with the information we now have of very recently emerged benches in Western Australia, at 2–3 and 5–6 feet above low-water mark, it would be understandable if these were confused with contemporary benches in an area of great tidal range. Recent detailed work on the black beach-sand deposits of southern Queensland by Beasley (1947, 1948) shows definite evidence of Recent sea levels at 10 and 5 feet above the present one.

Most observers agree that there is a well-marked bench at about 5 feet above low-water mark. Steers's "Lower Bench,"

in the sector between Bowen and Cooktown, is about the level of high-water neaps, which average 5–7½ feet above datum, representing a change in sea level of at least 5 feet. A "Higher Bench" stands generally about 8 or 9 feet above the lower, which is about 15–17 feet above datum.

If we accept this evidence, it follows that the 5–6-foot and 10–11-foot benches of the west and north are equivalent to the 5–7- and 15–17-foot levels on the eastern seaboard. It is as yet uncertain whether these differences in elevations are functions of the exposure or of differences in tidal range or are related to a slow tectonic emergence of the eastern Cordillera which began in Tertiary time.

Emerged reefs and shingle conglomerates may be correlated with these levels and the surfaces of many reef-flats and "lower benches" expose the truncated remains of large coral colonies in the position of growth. These corals must have expanded upward during one of the Recent periods of higher sea level and been planed off by the erosion of the present or one of the slightly higher mid-Recent seas. Few fossil *Lithothamnion* appear to have been associated with them.

The calcareous benches, above the present one at or about low-tide level, are deeply pitted and gullied, indicating reduction to base level, except where the surface is, or lately has been, protected by a blanket of beach rock or by a sand cay. Large areas of the reefs show no living corals and appear to be the planed surfaces of older reefs.

The question of negroheads, or "niggerheads," arises in connection with a discussion of emerged reefs. These terms refer to large blocks of dead coral, coral conglomerate, reef limestone, or beach rock lying on the reef periphery. The bulk of these are thrown up there by storms and cyclones, "jetsam," according to Saville-Kent (1893). Their accumulation was called a "hurricane beach" by Hedley and Taylor (1908). Most observers adhere to the "jetsam" hypothesis, although Agassiz (1898) favored the idea of Jukes (1847), Dana (1849, 1872), and Semper (1881) that some of them were solution relics of emerged coral platforms.

In the East Indies there are no hurricanes, and Umbgrove (1931, 1947) stated that there are no negroheads there. Most of the Queensland negroheads occur in the hurricane belt. They are absent in the Torres Straits region, which is free from hurricanes (Mayer, 1918). Agassiz observed, however, that some of the Great Barrier Reef negroheads, such as those of Three Isles, are composed of beach rock, which can form only in an intertidal zone. They could not have originated below the outer reef rim and have been thrown up by hurricanes. Undercutting of mushroom-shaped rocks and undermining by solution (Umbgrove, 1947) explain their origin, in the words of Agassiz (1898, p. 114), "as monuments of the former extension of beach rock."

Agassiz interpreted the negroheads on Cairns Reef and on "E" Reef (South Turtle) as remnants of an eroded reef. On "F" Reef, he found a huge negrohead, 9 feet high, consisting of one solid mass of *Porites* in its original position of growth. Another example of the solution type of negrohead is an emerged reef, 12 miles southwest of Raine Island (Jukes, 1847). According to Agassiz, the old reef breaks up into a number of isolated negroheads, which are clearly erosion relics. Such long, flat-topped, and undercut segments, termed coral "horses," have been noted in the Maldive area (Gardiner,

1931*b*; Sewell, 1936*a*), and in Tahiti (Crossland, 1928*a*).

Some excellent examples of solution-type negroheads have been observed by the writer on the southwest reef of the Pelsart group of Houtman Abrolhos, far away from the hurricane belt. They consist of lagoon limestone, in flat-topped blocks containing shells now extinct in the area, and identical with the deposits exposed on emerged platforms in the near-by islands. The writer thus accepts Agassiz's interpretation for certain examples and Saville-Kent's for others.[8]

REEF GROWTH

ECOLOGIC CONTROL OF REEF CORALS

Some consideration of the factors controlling coral growth in the Australian region is necessary. First of all, a brief note on ecologic controls is given, as these biologic factors are primary.

Observations here confirm the statement by Vaughan (1919, 1923*a*) that flourishing communities are restricted to waters where the average temperature does not drop below 18° C. in the coldest month (fig. 6). In Australia their range thus extends from Rottnest (32° S.) on the west to Moreton Bay (27°30′ S.) on the east (Hedley, 1925*c*, *d*; Schott, 1933, 1935).

Slight variations from normal salinity are fatal to corals. The complete absence of corals at Hamelin Pool in Shark Bay is due to high salinity, 4.8 per cent in the summer. Excessive fresh water, notably on the tropical Queensland coast, is equally inhibitory (Mayor, 1924; Hedley, 1925*a*; Rainford, 1925; Crossland, 1928*b*).

Because sediment in suspension is generally high in the Great Barrier Reef lagoon in Queensland but low off the west coast, which is also generally rather poor in corals, the inhibitory effect of sediment may have been overrated. Its effect is more to restrict light penetration (Verwey, 1931), because corals not only are phototropic but also require adequate light for the growth of their symbiotic commensal algae, the Zooxanthellae (Yonge, 1940). This is the probable explanation of the shallow depth of coral growth in the Barrier Reef lagoon, where coral heads are today growing only from depths of 6–7 fa'homs (Agassiz, 1898). Yonge (1940) has observed maximum growth here at only 2½ fathoms. In the clearer waters near the edge of the shelf, however, active growth may reach 15–25 fathoms.

Another important matter is exposure. Most of the Great Barrier Reef area is exposed to onshore winds, the southeast trades, whereas in northwestern Australia the same winds are offshore in character. Thus coral larvae may tend to be carried onto shore in the first area, but away from shore in the second. Aeration may also be reduced in the latter by the same factor. The Barrier Reef lagoon (Orr, 1933) is poor in nutrients, but the arid shores of the northwest may be still poorer.

GEOMORPHOLOGIC AND STRUCTURAL CONTROLS

Both geomorphologic and structural (tectonic) controls may influence the shape and loci of reefs. The latter may be

[8] Certain authorities have tended to discredit much of Agassiz's coral-reef work; it seems not too late to emphasize that, with all his faults, he was still a remarkably keen observer (see also remarks by Gardiner [1930] after his work in the Agassiz Museum, Harvard). Dogmatic opinions, often held by both geologists and biologists on coral-reef problems, were often based on the most limited evidence and in complete ignorance of one another's science. Today we recognize both subsidence *and* solution in lagoons, sedimentation *and* light as ecologic controls, solution type *and* jetsam type negroheads—all topics of late-nineteenth-century controversy.

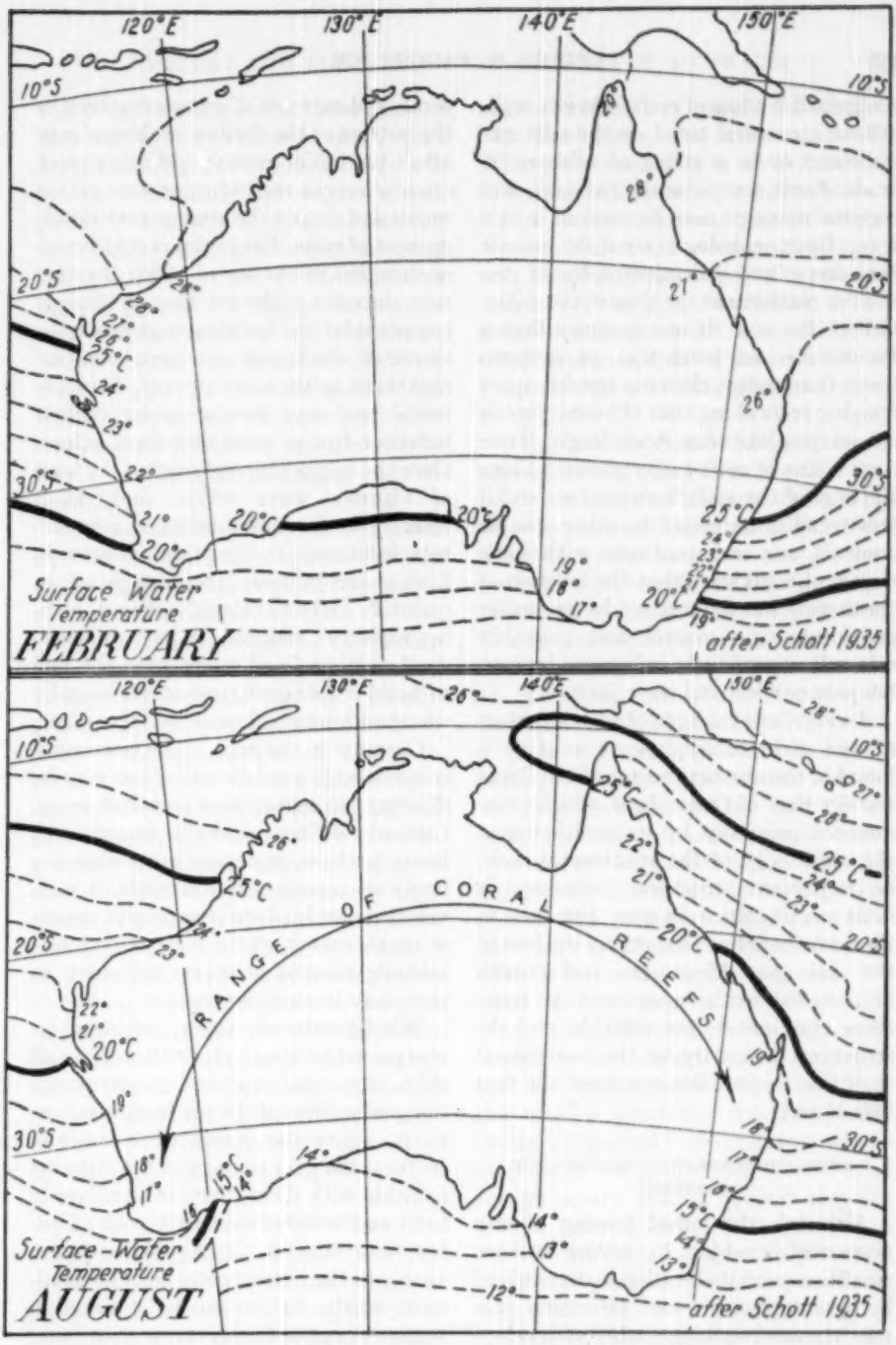


FIG. 6.—Maps to show distribution of surface-water temperatures around Australia for February (southern summer) and August (southern winter). Temperatures are in degrees centigrade (after Schott, 1935). The relationship of the range of coral reefs to the average temperature of 18° C. for the coldest month is indicated.

recognized if a line of reefs follows a well-defined structural trend on the adjacent mainland or in a string of offshore islands. Fault scarps, anticlinal axes, and regional upwarps may be marked in this way. Geomorphologic controls include the shape and paleogeography of the shelf or platform at the time of the initiation of the reef. If one assumes that a Pleistocene sea level was 40 fathoms lower than today, then one would expect fringing reefs along that old coast line in the warmer latitudes. Accordingly, if one finds a line of reefs today growing along the edge of the 40-fathom contour and if this trend is unrelated to other specific controls, one may postulate with some degree of confidence that the location of these reefs was determined by an earlier geomorphologic control and probably was only superficially influenced by contemporary wind and wave factors.

An excellent example of such structural *and* old geomorphologic control is found in the northern sector of the Great Barrier Reef of Queensland, which is influenced primarily by its position near the outer edge of the continental shelf, an important structural boundary (a fault scarp), which, in turn, lies close to the 40-fathom line that marks the foot of the main post-Pleistocene reef growth here. In the southern sector of the reefs, these two lines do not coincide, and the structural boundary of the continental shelf lies 20–30 miles out from the first belt of reefs.

WINDS AND CURRENTS AS MORPHOLOGIC CONTROLS

Although the initial footing of any given reef is subject to existing ecologic conditions and its location is determined by geomorphologic and structural factors, the most striking control of development and form of coral reefs is exerted by winds and currents. Currents reaching to the bottom of the shallow-shelf seas may affect the initial growth and subsequent orientation of a reef, whereas wave action molds and shapes the later growth development of reefs. The initial reef forms an obstruction in the lee of which the current deposits sediment that in time is colonized by reef-building organisms, the larvae of which may also have been carried there by the same current. Thus the initial reef may develop under current influence free of wind and wave action. Once the upgrowing reef reaches the level of vigorous wave action, and, as it emerges in the intertidal zone, new factors influence its direction of growth. Living reef-builders are broken off in quantity (only to rebuild with astonishing rapidity), reduced, in part, to sand-sized particles, and swept over the reef to build a sediment cone on its lee side, where additional colonization takes place.

Currents in the open ocean commonly coincide with wind direction but may be divergent in narrow seas and shelf areas. Currents do not generally exceed $\frac{1}{2}$-$1\frac{1}{2}$ knots in the open oceans but attain 2–5 knots in narrow seas. Currents of such velocities are ineffective as erosive agents of corals except where they carry a formidable amount of coarse sediment, as they may in narrow passages.

Wind-generated waves, impeded or obstructed by a reef, erode the reef on all sides, although the several quadrants are unequally affected. In the trade-wind regions, where the winds are essentially uniform the year round and the currents coincide with them, there is a uniformly built and oriented system of reefs (Hedley and Taylor, 1908; Wood-Jones, 1910). In the intertropical belt the dominant winds, the monsoons, blow each season in opposite directions (Gardiner, 1903; Krempf, 1927, 1930; Umbgrove,

1947). In the intermediate belts there are varying degrees of reversal of the trade directions and the monsoon directions. In the Australian shelf region a completely balanced wind reversal does not occur, as in the Java Sea (Umbgrove, 1947), but along the northern shelf areas there is generally a change from southeast to northwest winds for several months of the year (Teichert and Fairbridge, 1948). In the narrow shelf seas, where many of our reefs grow, the currents commonly do not coincide with the winds, so that many shapes of reefs are possible.

In most cases not only differing wind and current conditions operate but also the initial geomorphologic and structural controls all combine to create the final reef form. Thus the outer-barrier reefs of Queensland, already mentioned as reflecting these initial controls, are also greatly modified by the winds and currents. Although the over-all pattern of these ribbon reefs is dominated by a structural line, each individual ribbonlike segment is rounded off by the ingrowing horns that are the result of waves generated by southeasterly trade winds sweeping in through the narrow passages. These waves are retarded along their margins, resulting in turning of the wave front at right angles or even in an opposite direction to the initiating force. The distribution of coral larvae along the barrier edge is, furthermore, assisted by the regular north-south "Eastern Australian," or "Notonectian" current of Hedley (1910) (Paradice, 1925).

SHAPE OF PLATFORM REEFS

Examination of the shapes of platform reefs, as, for example, the inner reefs of the Queensland area, excluding barrier or fringing reefs, discloses certain characteristics apparently related to the prevailing wind direction. In the shallow Great Barrier Reef lagoon the current generally follows the southeast to northwest wind direction but becomes locally deflected by the reefs and in some places disturbed by powerful tidal streams (Fairbridge and Teichert, 1948, p. 70). The wind influence on reef shape is clearest in this area because the wind is southeasterly for about ten months of the year. Several stages in the reef development may be recognized.

Stage 1 is the initial reef patch, an irregular mound rising from the shelf floor (fig. 7). As soon as it breaks the surface, wave erosion will begin, and debris from it will be washed over to accumulate in a sedimentary cone on the lee side. When this bank becomes sufficiently shallow, further colonization occurs, and coral heads grow up, to be welded eventually to the initial reef. Examples are Ee Reef (15°29′ S.) and Cee Reef (15°32′ S.).

Stage 2 is reached when the initial platform develops two hornlike cones of sediment on either lee side and the exposed reef becomes crescentic or horseshoe-shaped. At this stage a definite differentiation in the morphological trend occurs. The smaller reefs remain compact, with small recurving cones of sediment, whereas the larger ones, over 2–3 miles across, acquire long, sweeping horns. This differentiation appears to be related to wave deflection, best seen from the air photographs. The horns of the smaller platforms are completely "wrapped around," or confluent, whereas those of the larger horseshoes are far enough apart that a lagoon may be formed between them. Thus the horns are built up from both ends and extend leeward in parallel. An example of this type is Hastings Reef (16°30′ S.).

The following stages of the larger and smaller types are considered separately.

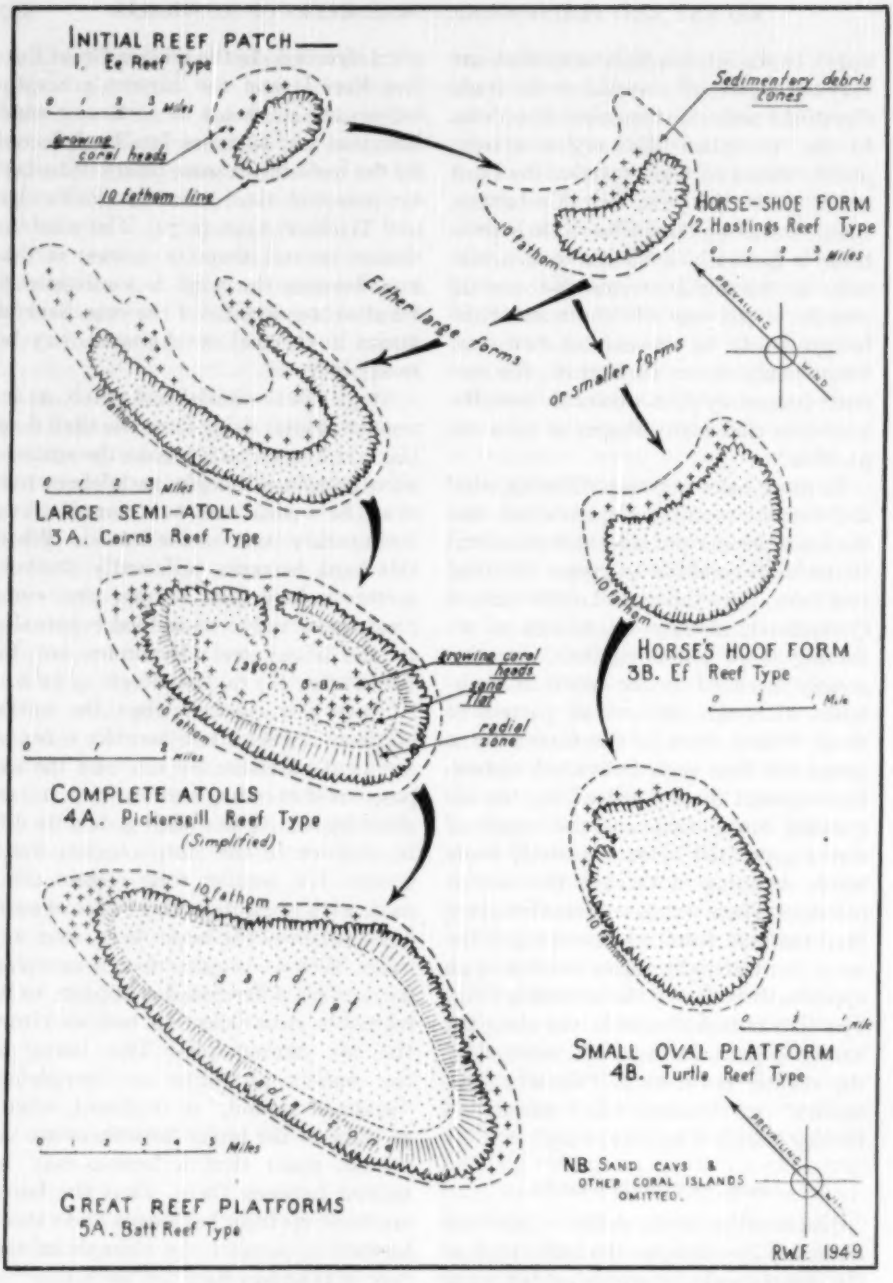


FIG. 7.—Evolution of shelf atolls and platforms from examples on the Great Barrier Reef (Queensland) shelf

Stage 3*A* is the larger type developed into a semi-atoll or almost atoll-shaped reef, with horns extending as far as 10–12 miles. The horn to the southwest generally grows faster because of the greater exposure of the south-facing reefs in the Great Barrier Reef lagoon, whereas the eastern one is generally somewhat protected in the lee of the outer barrier. The head of the small lagoon slowly fills with sediment, on which grow coral heads. Cairns Reef is typical of this stage (Hedley and Taylor, 1908) (pl. 5, *C*).

Stage 4*A* is the semi-atoll developed to a complete atoll. Small lagoons are quickly filled with debris washed in from the margins. Coral heads, growing up from the shallow floors, 6–8 fathoms deep, further block the inside. Pickersgill Reef is an excellent example of this type (fig. 4).

Stage 5*A* is the end-point in the evolution of the atoll developed from the larger type. The lagoon becomes gradually blocked by coral heads, filled in between them with "coral" sands, debris, Foraminifera, and precipitated calcareous muds. Finally, it becomes a sand-flat in which there are generally a few shallow pools left. In aerial photographs of the larger reef platforms the distinction between the former reef rim (the radial zone) and the featureless or mottled area of the sand-flat is striking. It appears that the great reef platforms, at least 10 miles in diameter, originated in this way. Batt Reef is a typical example (pl. 1, *B*).

Tracing next the evolution of the smaller forms following the initial horseshoe stage, we find the following stages.

Stage 3*B* is a horsehoof-shaped reef, formed by the gradual filling-in of the horseshoe or crescentic form. Such reefs are generally only a mile or two in diameter. Ef Reef (15°10′ S.) is a typical example.

Stage 4*B* is the end-stage at which the horsehoof has become elongated into a small oval reef. Lengthening to a sausage shape may occur through building of the debris cone to the lee. Aitch (H) Reef (14°55′ S.) and Turtle Reef (15°25′ S.) are examples (pl. 5, *B*). Many of these reefs are of considerable age, as shown by mid-Recent benches, and carry islands. At Low Isles, borings on the central reef-flat revealed that at least a certain portion of the central flat constitutes an infilled lagoon because the sediments beneath the surface of cemented debris consist of loose sand and soft mud and are composed locally of nearly one-third terrigenous clastics (Marshall and Orr, 1931).

In attempting to interpret these reef stages, it should be borne in mind that on the Queensland shelf area a relative degree of stability has persisted probably throughout the Recent epoch, at least during the last four thousand years. Furthermore, the sea level remained nearly constant, having dropped only 10 feet over this period. Accordingly, the developments in surficial reef morphology during the last four thousand years cannot be related to major changes of level of land or sea. The evolution in reef morphology can thus be related to the molding action of the wind, waves, and currents in the manner outlined above.

REVIEW AND COMPARISON

These conclusions, supported by many new data, confirm earlier views. In general, the recognition of the role of currents in controlling the shape of reefs goes back to Darwin (1842), Semper (1881), and Voeltzkow (1901, 1907). In the shelf reefs of the Java Sea and southwestern Celebes, Umbgrove (1929, 1930*b*, *c*, 1947) demonstrated the effect of currents in shaping the initial reef forms and the control of monsoon winds over the reefs and their islands.

On the adjacent shelf, off Indo-China, Krempf (1927, 1930) drew attention to the effect of the six-month alternation of the northeast monsoon and the southwest monsoon. Instead of a single horseshoe-shaped reef, known to form under the influence of a single wind direction, two contraposed horseshoes develop to form an atoll. Krempf applied this explanation to all oceanic atolls, without considering the circular volcanic foundations of some of the latter. We agree that in regions of alternating monsoons, many of the faros in the compound atolls on extensive banks, such as the Maldives, probably originated in this way and not by internal solution, as interpreted by Murray (1880, 1891), Agassiz (1903), and Gardiner (1903).

Wood-Jones, following his study of the Cocos-Keeling atolls (1910), concluded that the form of oceanic atolls was "entirely dependent upon the prevalent wind and seas." Although we cannot follow this view for all the oceanic atolls, it fits the shelf atolls best. Vaughan (1914*a*) explained similarly the shelf atolls of the Florida reef tract.

In the Great Barrier Reef area, Hedley and Taylor (1910*a*) emphasized the influence of the southeast trade winds on the morphologic development of the reefs. They summarize the process as follows (p. 413):

> The growth of an individual reef is shown to proceed in a regular cycle. If the reef reaches the surface with its axis along the wind, then its shape endures; but if across the wind then its extremities are produced backward, forming first a crescent, later a horse-shoe, and lately an oval, thus enclosing a lagoon. . . . In quiescence the lagoon walls broaden, the lagoon is obliterated with sediment, a vegetated sandbank spreads on the summit, and the atoll, grown to a cay, has arrived at maturity.

Certain of the oceanic atoll lagoons, particularly the smaller ones, are also filling up in this way, for example, Addu Atoll in the Maldives (Gardiner, 1903; Sewell, 1936) and many in the Pacific (David, Halligan, and Finck, 1904) and the East Indies (Umbgrove, 1939, 1947). However, the floors of many of the broader atolls with deep channels remain swept clean by currents. The rocky floors exposed at various depths, commonly at about 35–40 fathoms, probably correspond to the surfaces of intertidal induration of Pleistocene low-sea levels (such as Würm II). For this reason, Gardiner (1903, 1906, 1936) remained convinced that lagoons, in general, were not filling up. Nugent (1946) has demonstrated that only minor sediment filling has taken place in the larger Pacific atolls since the Pleistocene.

The observations of the British Museum Expedition of 1928–1929 in the Queensland Barrier Reef area favored the wind and wave explanation of Hedley and Taylor (Steers, 1929; Spender, 1930; Yonge, 1940). Stephenson, Tandy, and Spender (1931, fig. 3) depict the wave pattern of the southeast trade wind around Low Isles as traced from aerial photographs. The shepherding effect of the waves curving around the horseshoe-shaped reef tends to build up coral shingle all around the reef, although predominantly on the windward side. Thus a near-atoll form can be built without any wind reversal. The final rounding-off of the lee side of some of the big oval platform reefs may be achieved during the few weeks when hurricanes or the northwest monsoon temporarily reach down the Queensland coast. Their effects on Low Isles have been reported by Moorhouse (1936) and by Fairbridge and Teichert (1947, 1948).

The effects of currents and winds on the reefs of the Sahul Shelf have been studied by Teichert and Fairbridge

(1948). Monsoon reversal affects the reefs, although the orientation of some of them is controlled by the structural trends of the shelf basement. Rounded atolls are found near the outer shelf edge. A single wind direction is also found in the Abrolhos Islands, where a V-shaped shelf atoll in the Pelsart group, with a partly closed northern side, points to the south into the prevailing wind (Fairbridge, 1948*a*).

ORIGIN OF CORAL ISLANDS

Most authors agree that sand cays on reef islands are products of present-day wind and wave action (Gardiner, 1903). Observations and records of the appearance and disappearance of sand cays confirm this view (Edgell, 1928*b*). In general, the sand cay has been defined (Marshall, 1931) as an island that takes form in consonance with the action of the sea and wind. If the reef lies transverse to the wind, an atoll-shaped island may result.

The inner and island reefs in the Great Barrier Reef support Spender's views (1930, p. 200). He writes: "A sand cay is a structure whose stability depends on the equilibrium of various opposing forces brought into play at the leeward extremity of a reef." It is generally stable under the southeast trade winds but is liable to rapid erosion during the northwest monsoon or during hurricanes. However, the cay is partly protected against such attacks by the formation of beach rock. Spender's idea of "balanced finality" (p. 290), however, is questionable. The equilibrium can easily be upset by eustatic changes of sea level and other disturbing factors (Fairbridge and Teichert, 1948).

Saville-Kent (1897) observed how a shingle cay could also be formed in this manner in the Abrolhos, and in the Great Barrier Reef lagoon it is generally found on the windward portion of the reef, where the shingle tends to form a "rampart," a term first used on the Cocos Islands (Wood-Jones, 1910, p. 255). The ramparts are actually shingle beach ridges which may eventually form an island.

In the Queensland island reefs, Hedley and Taylor (1908) concluded that these "clinker embankments" were wind- and wave-induced and were "a prime factor in the building of a coral islet." However, the formation and preservation of multiple ramparts or beach ridges requires some slight emergence. According to Spender (1930, p. 286), "They depend on the existence of a flat higher than an inner reef, abnormally high therefore. . . . The conditions experienced by either reef are identical." However, ramparts are absent on the inner reefs, although they are present on the island reefs. Only on the island reefs, however, are there emerged benches associated with older ramparts. Likewise in the Abrolhos, shingle beach ridges are preserved only in multiples on the older, emerged islands (Teichert, 1947; Fairbridge, 1948*a*).

Similar conclusions have been reached in other areas. In the western Indian Ocean (Aldabra, etc.), a negative strand displacement was suggested by Voeltzkow (1901, 1907), who, however, believed it to be tectonic rather than eustatic because of its apparent suddenness. Gardiner (1903, p. 171) also indicated the probability of a Recent widespread drop in sea level on the basis of his observations in the Maldives. In a later paper (1906, p. 454) Gardiner suggested a glacially controlled eustatic drop in sea level, reaffirming this opinion subsequently (1936). Sewell (1928, 1932, 1935) reached similar conclusions for the Indian Ocean islands in general.

In the Pacific, evidence of this same mid-Recent eustatic drop, causing emergence of coral islets, has been noticed since the days of Forster, Vancouver, and Peron, as shown by Dana (1872), Gardiner (1898), Daly (1920), Wentworth and Palmer (1925), Ladd and Hoffmeister (1927), Crossland (1928*a*, 1931), Stearns (1941), L. A. Cotton (1947), and Tester (1948). Stearns and Tester refer only to a 5-foot drop, but a literature review suggests that there are several stages in the Pacific, just as there are in the Indian Ocean (Fairbridge, 1950*a*).[9]

Considerable evidence is coming in to support Gutenberg's conclusion (1941) that a reversal, a rise in sea level, is now in progress (Fairbridge, 1947), and climatological and glaciological work by Ahlman (1949) fits closely into the picture. If this tendency continues for a longer period, we may expect to see the complete destruction of many low coral islands but a vigorous increase in coral growth.

DISTRIBUTION AND DEVELOPMENT OF REEF TYPES

If we were to assume that the present continental shelf was a stable feature and that the sea level remained eustatically constant, then we should be able to explain the distribution of the shelf reefs in ecologic terms. The distribution, however, cannot be wholly explained by ecology.

GREAT BARRIER REEF AREA

Fringing reefs.—In eastern Queensland the fringing reefs have been shown to increase steadily in size and frequency to the north. In the region south of Cooktown they are quite rare, except locally around continental offshore islands, whereas to the north they are almost continuous. Ecologic controls seem to explain these facts. In the south, heavy rains depress the salinity and introduce much sediment into the littoral and offshore belts. North of Cooktown there is less rainfall, and the sandstones of the mainland do not yield excessive mud. Slightly higher sections of the fringing reefs toward the interior ("promenades" of Steers) in many cases form a platform of dead coral on which this mud easily accumulates, an indication also of the age of the reef, i.e., it is not all contemporary. Eustatic effects may also be seen in slightly submerged sections of the reef offshore. A shelf of 1-fathom depth of this sort in the New Hebrides suggested to Baker (1925) that the fringing reef was a thin veneer. We would visualize the fringing reef rather as a succession of veneers, a laminated structure, consisting of successive layers of fairly rapidly grown coral separated by indurated bands which represent erosional surfaces during periodic emergences under eustatic control.

Island-reefs of the Inner Channel.—The inner part of the Great Barrier Reef lagoon is an almost reefless zone from the Torres Straits down to southern Queensland. This is the so-called "Inner" or "Steamer" Channel of Spender (1930, p. 277) or "Australia's Grand Canal" (Richards and Hedley, 1925) because it

[9] As a matter of interest, it was the effect of this Recent eustatic drop in sea level, causing such widespread areas of eroded (and "dead") reef surfaces, that caused the biologist Crossland mournfully to lament: "The age of corals is past" (1928*a*, p. 605). An apparently opposing view was expressed by Daly (1919), who referred to the rapid postglacial upward growth of the reefs and concluded that "the present vigorous growth of corals is probably exceptional," the preglacial development being distinctly poorer. Andrews (1902) previously realized this by contrasting the thin, bedded character of the Tertiary reefs with the thick, massive growths of the postglacial period. Actually, both Crossland and Daly were partly right. The very recent drop in sea level has killed off and planed down the more extensive flourishing reefs of some four thousand years ago.

furnishes an excellent protected steamship route. It is 10–20 miles in width and 15–20 fathoms in depth. The 15-fathom level is sufficiently constant to form a 15-fathom terrace. In places this terrace seems to grade gently toward the shore. However, there seem to be intermediate steps at about 6–8 fathoms and at about 3 fathoms.[10]

Reefs in the Inner Channel are restricted to the section between Cairns and Thursday Islands, where they number about two dozen and are nearly all of the small oval variety (our type 4*B*) crowned by islands (type 4). Most of them expose a small core of mid-Recent coral material and thus must have grown to their present form over four thousand years ago. Examination of the Admiralty charts reveals that they have certain features in common. They arise from the 15-fathom terrace or from a slightly higher bank. They are elongated north-south to north-northwest–south-southeast, the principal structural trend of the mainland as well as the main wind and current direction. Thus the basement may be either a submerged "continental" island or a sedimentary bank.

The living coral on these reefs does not extend to the floor of the shelf. At Low Isles there are soft, muddy sediments offshore, which are stirred up by storms. Living corals rarely extend more than 6–7 fathoms down the sides of the reef. Therefore, the reef could not have risen from the bottom under present conditions. It probably grew up during the postglacial eustatic rise of sea level.[11] According to Zeuner (1945), the last major glacial stillstand, corresponding to Würm III, some twenty-two thousand years ago, was at about 16½ fathoms, which approximates our 15-fathom terrace. Therefore, our reef may have formed as a fringing reef on a small rocky headland and have grown about 109 feet during the fairly rapid postglacial rise of the sea, the Flandrian transgression, to the level of the mid-Recent "high." As suggested by shallow borings, Low Isles may have passed through the stages of a small horsehoof reef and a near-atoll before filling in (Yonge, 1940).

Inner reefs.—The next zone in the Great Barrier Reef lagoon is that of the great platform reefs, the "Inner Reefs" of Jukes (1847). In the sector north of Cairns these occupy a belt 10–15 miles wide. Except for the deep cross-channels opposite the openings or passes in the outer barrier, these inner reefs block the central part of the lagoon almost completely.

The inner reefs between Cairns and Cape Bedford seem to have been initiated as short northeast-southwest reefs, which grew with two long arms in a northwesterly direction, developing in part into simple crescents, in others into near-atolls, and in yet others into infilled massive platforms. The latter stages are products of wind and current, but the initial northeast-southwest trend may have a structural origin because it is parallel to numerous cross-faults on the adjacent mainland.

In the sector between Cape Melville and Cape Direction, the reefs trend dominantly northeast-southwest, corresponding to the tidal streams (see Admiralty "Pilot," vol. 4), but also with the frac-

[10] The large-scale cross section of the shelf and reefs between Hinchinbrook Island (18°28′ S.) and Myrmidon Reef, 62 sea miles from the shore, prepared by the Hydrographic Department for Richards and Hedley (1925, p. 26), depicts shoals for 2 miles offshore with a sudden drop to 3 fathoms, our first terrace. It then slopes gently down, to flatten out again in 12–15 fathoms between 6 and 11 miles offshore. It drops gently to 20–25 fathoms, from which terrace rise the inner reefs, Bramble and Trunk.

[11] There is no trace of pre-Würm raised terraces.

ture trends of the mainland.[13] These northeast-southwest trends are repeated in the long reefs leading to the Flinders Entrance, through which pass powerful currents in both directions. The youthful volcanic islands of the Murray group lie in this area. Probably fractures of the shelf here deflect the currents and thus control reef growth (Yonge, 1940, p. 376).

In general, the inner reefs rise from about the 25-fathom contour, which seems to mark the edge of a somewhat lower terrace, analogous to the 15-fathom terrace described above. Exceptions may be explained by sedimentation in places and scouring elsewhere. Further study is needed.

As will be seen in the review of the deep borings, the two inner reefs, Michaelmas Cay and Heron Island, have both grown up from a foundation of soft sediments, but this may not apply to all inner reefs. In the Lizard–Direction Island area every transition may be seen between a high rocky island with a fringing reef, or even with an incipient barrier as at Lizard Island, a broad reef platform with a tiny projection of continental rock, and a reef platform. Although the latter is indistinguishable in morphology from reefs with proved foundations of soft sediment, its location on a shelf ridge, corresponding in trend with the continental islands, suggests a rocky basis.

The inner reefs south of Cairns are generally small, with extensive sediment trains dotted with coral heads. They are more juvenile than those in the northern sectors. They lack the protection of a continuous outer barrier. To the southeast the continental shelf broadens, and the inner reefs occur progressively farther from the mainland. Little is known of them because many of them have not been resurveyed since they were originally plotted by Flinders nearly a hundred and fifty years ago. As far south as Townsville they show similarities to the northern types.

Short crescentic inner reefs are found at about 17°–18° S. Paradice (1925) erroneously referred them to the Outer Barrier. On the outer side of each reef crescent is a narrow windward shelf at 2–3 fathoms, common on exposed reef margins, below which the reef drops steeply to the shelf floor at about 25 fathoms. This steep windward slope is mostly covered by sand and dead corals.

Howie Reef, in this area, shows distinct lower platforms. It is a large, windward crescent, 7 miles long, facing east-southeast, and dropping precipitously on the southeast to 25 fathoms. Behind the crescent there is a smooth platform of about 2–5 fathoms deep and 2 miles wide, studded with coral heads, many of which reach the surface. A steep scarp leads down to another platform at 11–17 fathoms (averaging 15 fathoms), which extends another 2 miles leeward. Its southern margin is bordered by a long spitlike projection of the 2–5-fathom terrace. A 5-fathom drop in sea level would produce a near-atoll lagoon, floored by the lower terrace. Outside this 15-fathom platform there is a final drop to the 25-fathom platform with a sediment train partly obscuring the scarp on the lee (fig. 8).

Southeast of Townsville (Magnetic Passage) the inner reefs become irregular and are bordered by a better-defined

[13] Richards and Hedley (1925, fig. 6) depict east-west fractures here, although they describe in detail the northeast-southwest structural trends of the Flinders group. Richards and Hedley recognize the diagonal tectonic patterns in the Cape York–Torres Straits area, where the reef platforms are aligned east-west, controlled by the powerful currents passing through the straits.

outer barrier. They become more distant from the mainland, and the average depth drops to about 35 fathoms.

South of the Capricorn Channel occur the reefs and islands of the Capricorn and Bunker groups, morphologically similar to the inner (platform) reef types but forming the outermost reefs of the shelf, although they lie some 30 miles within the continental border. They rise from about 25 fathoms along a distinctly linear trend running through Lady Elliot Island to Sandy Cape via Breaksea Spit. The latter consists of dead coral and sand at a depth of 1–3 fathoms. The over-all length of the Capricorn-Bunker group is about 150 miles. Between this group and the coast, about 40 miles away, are no

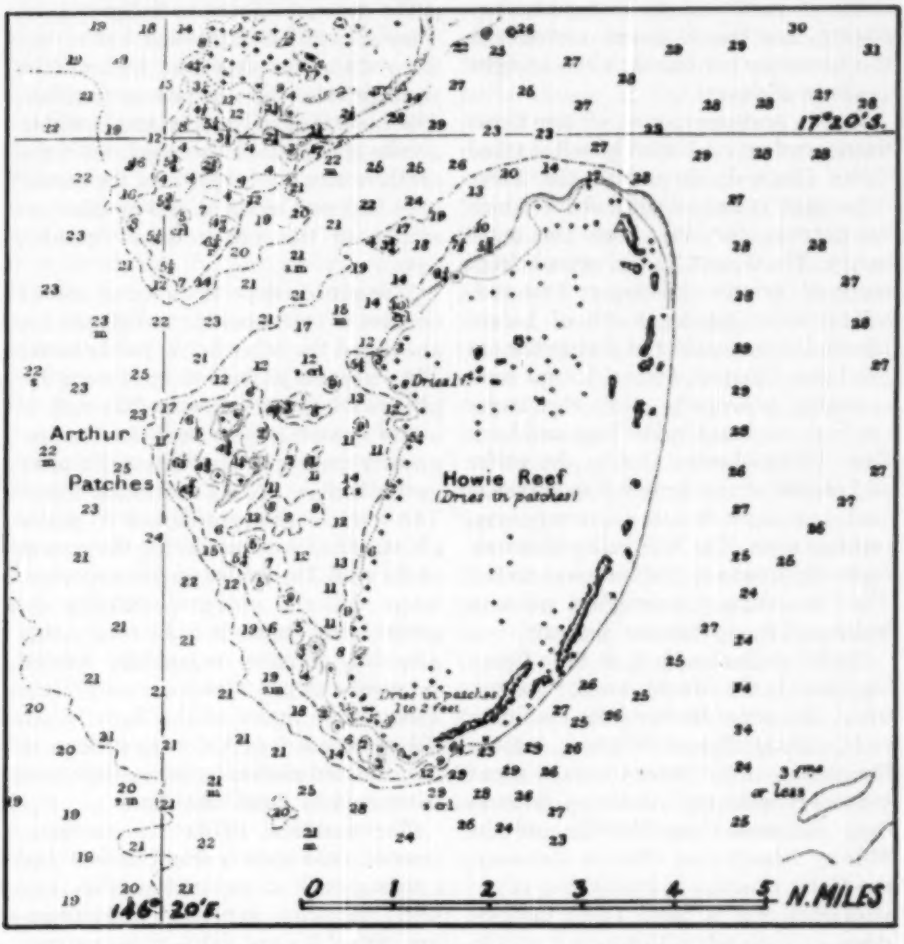


FIG. 8.—Chart of Howie Reef (from Admiralty Chart no. 2350), showing typical crescentic form with submerged platforms at 2–5 fathoms, 11–17 fathoms, and 25 fathoms, partially obscured to leeward (northwest) under recent sediment train.

coral platforms except two northwest-southeast–trending areas, one marked by Heron Island, Wistari Reef, Erskine Island, Mast Head Island, etc., and the other by North-West Island. To judge from the deep boring on Heron Island discussed below, they seem to be built on sedimentary trains, running in at right angles to the shelf edge. Most of the present reefs of this group are oriented northeast-southwest to east-west, suggesting that the dominant currents in this broad part of the shelf run at right angles to the coast.

In the northern sector of the Great Barrier reefs, there is a fairly well-marked Outer Channel, similar to the Inner ("Steamer") Channel noted above, which lies between the inner reefs and outer barrier. The Outer Channel extends from north of Trinity Opening to One-and-a-Half Mile Opening north of Lizard Island. It is not so clear of obstructions as the Inner Channel, several lateral reefs occurring principally near the larger openings, such as Cruiser Pass and Lark Pass. Flight observations by the writer and cruises of the British Expedition of 1928–1929 show that there are very long, reef-free areas. The Admirality Chart erroneously marks it "full of coral reefs." The clear channel averages 8 miles in width and 15–25 fathoms in depth.

North of One-and-a-Half Mile Opening, there is the "double-barrier" sector, where the outer barrier joins the inner reefs, cutting off a 40-mile-long, lagoon-like section of the Outer Channel, about 5–6 miles wide and about 15 fathoms deep. Between Cape Melville and the Murray Islands and Flinders Entrance, the Outer Channel is mostly free of obstructions and is some 15–25 fathoms deep.

South of Trinity Opening, there is no definite outer barrier and thus no determinable Outer Channel. Southeast of Magnetic Passage, off Townsville, however, the outer barrier is more definite, and there is a channel separating the outer from the inner reefs which appears to be badly obstructed in places by reef patches.

Outer-barrier reefs.—The outer-barrier reefs, as noted above, are divisible into two sectors. The northern-zone reefs are of the crescentic ribbon or linear variety. They are related to the edge of the continental shelf, as they lie a few hundred yards within the 100-fathom contour. The 1,000-fathom line in places is within 4 miles of the reef. This steep front of the northern barriers is typical of the barrier reefs of New Caledonia and of other examples in the mobile belts (Spender, 1930).

The outside slope of the outer barrier appears to be steepest in two places, one above and the other below 100 fathoms. The steepness of the outer-reef slope implies rapid upward growth of the reef, as a slow upward growth would lead to progressive erosion and formation of a more gentle incline (Von Lendenfeld, 1902). The terraces thus correspond to major stillstands of sea level during the growth of the reef. The double step is not everywhere uniform, owing to differing degrees of subsidence. It is, however, characteristic of many organically formed submarine slopes (Gardiner, 1915). It is absent from many of the East Indian slopes (Kuenen, 1933), owing perhaps to more limited erosion during stillstands or to excessively rapid subsidence.

The southern, unlike the northern, zone of outer reefs is discontinuous and consists only of isolated patches and small platforms. Between Trinity Opening (16°30′ S.) and 18° S. there are only submerged ridges (Paradice, 1925). The landward shelf floor is generally 40

fathoms, exceptionally deep for the Queensland shelf. There is an outer ridge at about 20–30 fathoms, although locally it forms an outer shoal of only 3–5 fathoms. On the charts it is marked "coral" or "shingle," but there are no data to show whether the coral is living or dead. The seaward side of the ridge falls steeply to 40 fathoms, where there is generally a narrow shelf, and then descends down to the 100-fathom line. This outer ridge may be a drowned outer barrier. Reef-building organisms apparently grew up quickly from 100 to 40 fathoms. A long erosional interval may have formed the 40-fathom platform during the early Würm, and, after further recession of the sea, fringing reefs colonized the old foundation, and a subsequent rapid rise of sea level produced the steep-sided 20–30-fathom reef. A very rapid subsidence may have followed, drowning the barrier save for a few places.

If the borings at Michaelmas Cay and Heron Island can be taken as a guide, the massive coral framework may go down to 100 fathoms. The narrow, shallow seaward shelf may be an abraded platform, possibly Würm II in age,[12] and the shallower depths on the landward side would be a composite of a later platform and accumulated sediment.

Thus the outer southern reefs were probably initiated as fringing reefs in early Würm times, whereas the northern ones, although similar in type, were probably aided by a still deeper basement of earlier Pleistocene, or even late Tertiary, origin. In both, the bases of the reef structures are probably well below the lowest limit of Pleistocene eustatic lowering, and consequently subsidence must be assumed.

[12] According to Zeuner (1945), the sea level of Würm II was probably about 38½ fathoms, or 70 meters, below that of today.

Certain reefs in the Coral Sea provide further evidence of subsidence. The map (fig. 10), based on Admiralty charts 2763 and 2764 and Spender's contoured sketch (1930, p. 274), shows the northern sector of the barrier reefs to be bounded by the 1,000-fathom contour. North of Trinity Opening, this contour swings north again, encircling Bougainville Reef and the Osprey Reef atoll, and then swings southeast past Saumarez reefs, to come close to the coast again near Sandy Cape. Following the pattern of this contour fairly closely is the 600-fathom line, within which rise a whole series of atolls, partly drowned atolls, and near-atoll reefs (Flinders Reefs, Diane Bank, Willis Island, Lihou Reefs, Marion Reef, and Saumarez Reefs). This great bank is shallowest in the south, 200 fathoms around Saumarez, 300 fathoms around Marion, and is tilted gently north, so that Flinders and Diane reefs rise from about 400–500 fathoms, Bougainville rises from over 600 fathoms, and Osprey from 700–1,400 fathoms.

Paleogeographic and geomorphic evidence supports the theory that this tilted bank is part of the ancient land mass of "Tasmantis," which subsided farther than the present continental shelf. The growth of atolls on it supports a Tertiary-Quaternary subsidence. Hedley and Taylor (1908, p. 413) remarked: "Whatever the history of the Great Barrier Reefs, the reefs of the Coral Sea, such as the Lihou Reefs, Flinders Reefs, and the Herald Cays, shared it." The structural and sedimentational problems in this connection are considered separately below.

SAHUL AND NORTH-WEST SHELF AREA

We turn now to the distribution and development of reefs in the north and northwest (fig. 9). Fringing reefs, except-

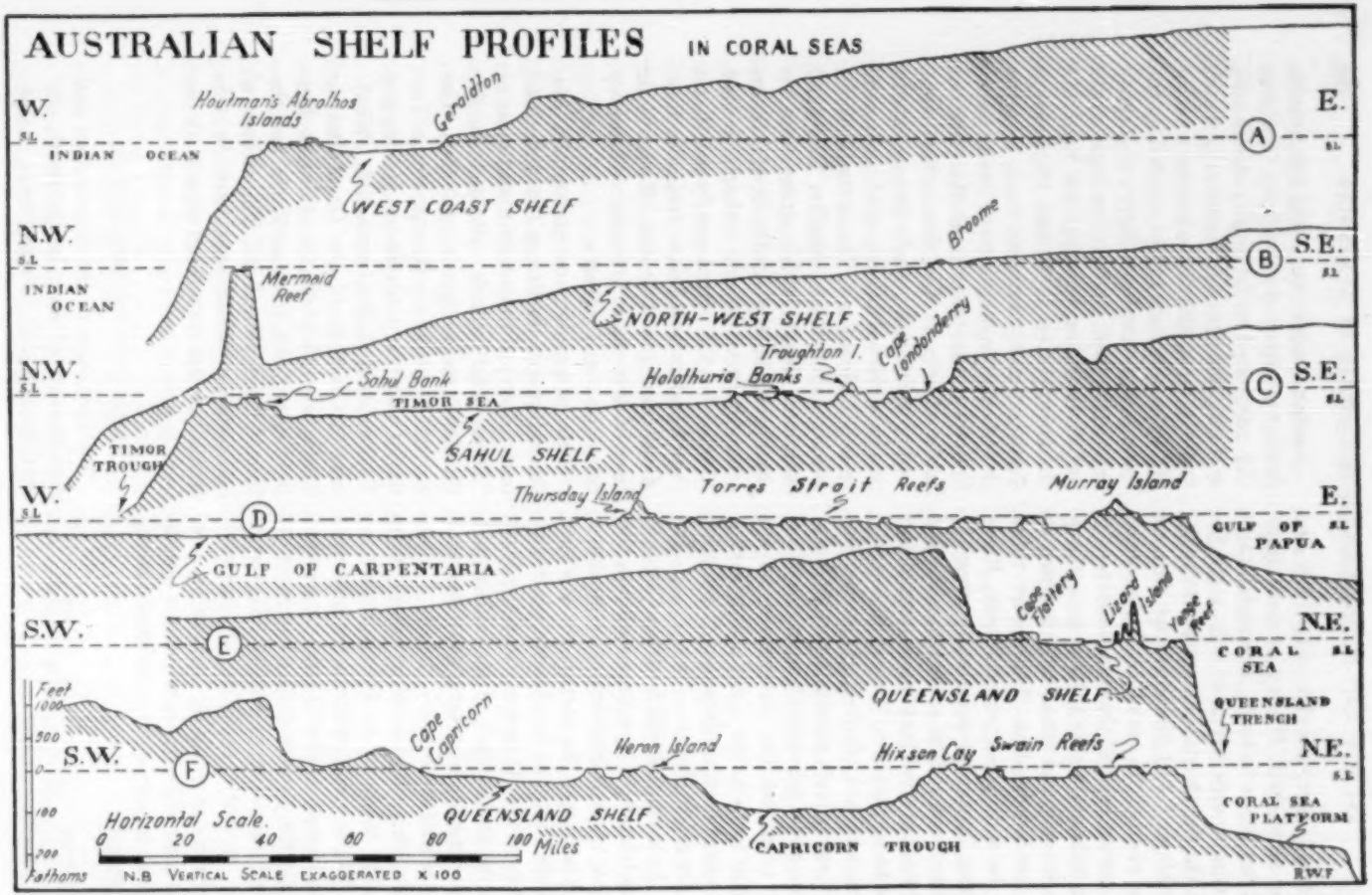


FIG. 9.—Australian shelf profiles in coral seas. Notable points include (*a*) the extremely variable width of the shelf (probably due *not* to variation in wave erosion or rock hardness but to structural controls); (*b*) the relative flatness of the shelf, apart from the irregularities due to coral reefs and a few "continental" islands (suggesting wave planation under varying sea levels); (*c*) the shallowness of the shelf, except on the margins the North-West Shelf and Sahul Shelf and the southern Queensland Shelf (where active subsidence appears likely).

ing those around offshore islands, are missing from the Gulf of Carpentaria and from the mud-filled bays, Van Diemen Gulf, Joseph Bonaparte Gulf, and King Sound, of the Northern Territory and North-West area. Along the tropical north coast, however, they are extensive. The sand-strewn sector of the Eighty Mile Beach is devoid of shore reefs, but fringing reefs flourish from Port Hedland to the North-West Cape. Their distribution is ecologically controlled.

Near-shore reef platforms are common in the shallow, current-swept Clarence Straits, between Melville Island and the mainland near Darwin, the broad Holothuria Banks and associated reefs of the North Kimberley, across the mouth of Collier Bay and King Sound, Adele Island and the Buccaneer Archipelago reefs, west of Cape Leveque, in the Lacepedes, and on the shallows off the coast between Port Hedland and Onslow. Almost all are in part structurally controlled because they rise from structural ridges or broad upwarps along ancient geological trends. They are shallow-water in character and do not suggest subsidence. Some upgrowth during the post-Pleistocene rise of sea level is probable, in view of the precise delineation of structural features in some cases, as Adele Island (Teichert and Fairbridge, 1948).

There is no true barrier reef along the margins of the Sahul or North-West Shelf areas, although Molengraaff (1930, p. 70) refers to it as a "much interrupted barrier-reef." On the Sahul Shelf opposite Timor, the margin is marked by a series of isolated ovoid reef platforms, such as Ashmore and Fantome, and a number of "drowned" coral banks, such as Sahul Bank. To the southwest the line is marked by a number of isolated atolls: Seringapatam, the paired atolls of Scott Reef, and, on the North-West Shelf, the three Rowley Shoals, Mermaid, Clerke, and Imperieuse reefs.

These shelf-margin reefs rise precipitously from depths ranging from about 100 to over 300 fathoms (Scott Reef, Chart AUS. 094). Thus they rise from an abnormally depressed shelf. Soundings suggest that there may be a quickly grown column of coral beneath, say, Scott Reef, not less than 1,800 feet thick. There is no evidence of Recent or Pleistocene volcanoes in this area, although the possibility of volcanic foundations for these atolls is not entirely out of the question (Teichert and Fairbridge, 1948, p. 246).[14]

The bathymetric data on the shelf and the geological history of the adjoining mainland suggest that these atolls lie in two adjacent areas of crustal subsidence, Seringapatam and Scott in one, opposite the drowned northwest Kimberley coast, and Rowley Shoals in the other, opposite the Desert Basin, a long active, paralic subsidence (Fairbridge, 1950*d*). The Ashmore-Sahul reefs rise from deep water, again too deep for reef growth. Their position opposite the mobile Timor Arc and its fronting trough seems to imply subsidence. Drowned reefs, like the Sahul Banks, suggest excessively rapid submergence.

A number of scattered reefs, like Vulcan, Heywood, and Browse, seem to rise from the edge of an inner shelf which is marked by an escarpment at 55–60 fath-

[14] The atoll reefs of the Sahul and North-West shelves are all nearly oval in plan, with no trace of the scalloped outline commonly found in the larger Pacific oceanic atolls, believed to rest on volcanic stumps. Their scalloped outline may be due to submarine slides down the volcanic cone (Fairbridge, 1950*e*). The true oval form is generally restricted to small juvenile reefs or to nonvolcanic shelf atolls. Its occurrence in large atolls off northwestern Australia may indicate that they are really shelf atolls and not volcanic-based structures.

oms (Teichert and Fairbridge, 1948). This step is fairly sharp, but sinuous in plan, and is probably an erosional feature. In elevation it corresponds closely to the level of the Würm I emergence. It may be postulated that these intermediate platforms are of late Pleistocene origin and grew up from very isolated fringing reefs or shallow-water reef patches during the last eustatic rises of sea level. Active tectonic subsidence may have been involved, but the erosional character of the inner shelf edge suggests that this submergence was recent and mainly eustatic.

This explanation is insufficient for the outer row of reefs. Here we must postulate tectonic subsidence, related partly to Quaternary down-warping of the northeast Indian Ocean and associated disturbance in the adjacent East Indian orogenic belts. The general youthfulness of subsidences and the formation of marine basins in this part of the Indian Ocean and the adjacent archipelago have recently been recognized by numerous workers and reviewed by the writer (1948*c*).

WEST-COAST SHELF AREA

As previously indicated, the bulk of the west-coast reefs are calcareous sandstone reefs, with veneers and fringes of living coral. These sandstone reefs occur all around the coasts of western and southern Australia and are truncated benches in Pleistocene eolianite rocks, which were formed as coastal sand dunes in parallel ridges along the various low-sea shores of the glacial periods. Offshore reefs of this type have, in places, the character of single or even multiple pseudo-barrier reefs. Their evolution as part coral reefs is a secondary and quite recent event. Coral-reef patches in this area are likewise Recent and of shallow-water character. Coral-reef veneers on eolianite bases occur as far south as Rottnest Island (32° S.) and farther south in raised reefs of the warm Pleistocent interglacials. There is no evidence of tectonic influence.

The Houtman Abrolhos Islands alone are purely coral in character. These rise from 25 fathoms near the outer edge of the continental shelf. The older core of coral-reef rock and massive lagoon limestone suggests a Pleistocene or older age. This core is truncated by late Pleistocene and mid-Recent erosion and veneered by younger coral reefs.

All the Abrolhos reefs and associated sediments may be explained by the development of a series of normal shelf atolls during or prior to the Pleistocene. These emerged during late glacial, low eustatic sea levels and were deeply dissected by karst erosion. With the ensuing stages of successively higher sea levels, they were truncated and beveled by marine erosion, which was accompanied by further reef growth fringing the old cores (Teichert, 1947; Fairbridge, 1948*a*).

No subsidence or tectonic influence is evident. The isolated position of the Abrolhos reefs remains unexplained. From consideration of the fault-block tectonics of the adjacent mainland, it seems possible that the Abrolhos lie on the extension of one of the tilted blocks, bordered by Hill River fault, which may have exposed a core of pre-Cambrian granite that offered a better footing for coral growth than did the sands and calcareous sandstones prevalent along this coast.

FOUNDATIONS OF SHELF REEFS

The three-dimensional, rather than geomorphic, aspect of reefs is of the greatest interest to the stratigrapher. The interior constitution of the reef

stock, its relation to sedimentation on the shelf and in the lagoons and to the hard-rock basement are of special interest. Some of the most illuminating information comes from borings, but, because the shelves are mostly covered with contemporary and comparatively recent soft sediments, it is appropriate to consider the question of such sedimentation first. The borings will be discussed below.

CONTEMPORARY SEDIMENTATION

Understanding of reefs is incomplete without some knowledge of the sediments in and around them. No extensive survey of contemporary sedimentation on the shelf in the Australian region has been made, and only scattered observations are available. The Admiralty charts carry references to the sediments, but the information is limited to a general classification of sand, shingle, mud, coral, etc. The best-known area is that of the Great Barrier Reef.

Great Barrier Reef area.—Rapid accumulation of sediment is now in progress in the Queensland barrier-reef area. The sediments are of two main types, namely, those coming from a well-watered and mountainous mainland and those calcareous sediments derived from the coral reefs. Chemically precipitated calcareous muds are of minor importance.

A rough idea of the distribution of the sediments may be obtained from the Admiralty charts. Bryan (1928) mentions a large number of references to sands and coarser clastics near the outer edge of the shelf. These may be carried over from the outer barrier reefs, but, where the latter are absent, they may represent residuals of low Pleistocene seashores, a feature common to many other shelf margins.

The terrigenous material extends over the entire shelf and is distributed in part by the strong currents of the barrier lagoon. At times during the wet season the lagoon waters show dark-brown discoloration. Large mudbanks at the mouths of many of the rivers are continually growing. The organogenic material is likewise mainly clastic, derived from the reef communities through wave action. Foraminifera make up one-sixth to one-half of the constituents in the average "coral" sand.

A limited amount of calcium carbonate mud is precipitated from the sea water in the reef-flat pools and lagoons under certain physicochemical and possibly biochemical conditions (Black, 1933; Hatch, Rastall, and Black, 1938; Smith, 1940). This material is ultimately derived from exposed coral by chemical solution in sea water under the conditions of the intertidal zone (Fairbridge, 1948*a*, 1950*c*).

Nearest the land the terrigenous material commonly exceeds 90 per cent, whereas in the vicinity of a reef the organogenic clastic materials and chemically precipitated calcareous muds may form 98 per cent of the total. The effect of currents in the vicinity of reefs is striking. A long cone of coralline sediment is generally found on the northwest side of many reefs, whereas the steep southeast, northeast, and southwest sides are swept clear of sediment. Deep channels skirt many of the reefs, suggesting current scour of the shelf floor.

Massive coral and coralline algae form the nucleus or framework of reefs, in and around which the clastic material is heaped. However, the reef-builders form an extraordinarily small part of the reef stock. The reefs, in turn, occupy an extremely small part of the continental shelf. The proportion of massive reef rock (grown *in situ*) to the total sediment on the shelf is thus very small—a factor of considerable significance for the stratig-

rapher, looking for evidence of reefs in ancient rocks of limited outcrop.

Although isolated corals may grow locally in favored spots on the shelf floor from 20 or even 10 fathoms, the bulk of the floor is coral-free. The growth of pinnacles and mushroom-like coral heads takes place only in favored localities, starting at 6–7 fathoms in the inner parts of the lagoon, as noted by Agassiz (1898). In the clearer water behind the outer barrier, coral-head pinnacles may, however, grow up from 15 to 20 fathoms, although it is doubtful that the bases are alive (Paradice, 1925).

In many parts of the Great Barrier Reef lagoon very rapid accumulation of material may be taking place, and, as noted above, it seems that most of the big reef platforms formerly had central lagoons which have been gradually filled in, as shown by Hedley and Taylor (1908) on Cairns Reef and by Stubbings (1938) in the lagoons of Lady Musgrave and Pickersgill reefs.

At Low Isles, Marshall and Orr (1931) tested contemporary sedimentation by setting catchment jars at various points on the reef. The calcareous material of the "coral" sand generally consisted of about one-third coral debris, one-third Foraminifera, and one-third molluskan and other calcareous debris. The proportion of calcareous material to nonsoluble varied from 98:2 in a protected, to 58:42 in an exposed, position after a heavy storm, showing how much terrigenous material is scoured up from the shelf floor during heavy weather. Hand borings on the reef-flat showed, in addition to reef rock and coral debris, soft gray muds at shallow depths which consist of 59 per cent acid-solubles, suggesting a former lagoon filled in by sediment during storms. According to Marshall and Orr, the Great Barrier Reef lagoon is at least 33 per cent more turbid than the English Channel. The sea floor around the reef consists of soft gray mud in depths of 10–15 fathoms, except to leeward, where a coral sand cone slopes down from 5 fathoms just off the reef to the general level of the lagoon.

Samples from other islands of the Barrier Reef, reefs, and lagoons have been reported on by Stubbings (1938). Of the acid-insolubles, only quartz was recognized, and that up to a maximum of 5 per cent in the Lady Musgrave lagoon. Among the calcareous materials, molluskan fragments were generally dominant, followed by either coral or foraminiferal material. Minor constituents included Porifera, Polychaeta, Alcyonaria, echinoids, Crustacea, *Halimeda*, and coralline algae. The rarity of coralline algae here is surprising, as they are elsewhere very common. As all the samples came from island reefs or inner reefs with sand cays, they were hardly a representative cross section of the reefs.

An analysis of reef-flat samples from Murray Island by Vaughan *et al.* (1918) showed that Foraminifera and calcareous algae play a dominant role, whereas Coccolithoporidae, alcyonarian spicules, and diatoms are subordinate or rare. Very little of the clay-sized fraction is preserved on the reef-flat; and, if $CaCO_3$ is precipitated here, it is probably swept off into deep water.

Much has been written about the limiting effects by sediment on coral growth. Le Conte (1856), in his explanation of the Florida keys, suggested that a barrier reef might be the same as a fringing reef, but "limited on one side by the muddiness of the water, and on the other by depth." Guppy (1884, 1887), in explanation of the barrier reefs in Bougainville Straits, Solomon Islands, came to the same conclusion.

Gardiner regarded submarine solution, the now rejected theory of Murray (1880, 1891) and Agassiz (1903), and others, as mainly responsible for the formation of lagoons, but he recognized the subsidiary action of sediments in preventing the growth of coral on the lagoon floor (1903, p. 334). Later he commented specifically on the chemical precipitation of $CaCO_3$ on lagoon floors (1930, 1931*a*). In the mangrove swamp-lined lagoons, such as Aldabra, on the other hand, there is much swamp acid and therefore increased solution of $CaCO_3$ along the raised limestone margins. This is later precipitated in the lagoon, where coral growth is entirely inhibited (Fryer, 1910). White precipitated $CaCO_3$ mud is found in most lagoons (Fairbridge, 1948*a*, p. 27).

In the Cocos atoll lagoon, Darwin (1842) pointed out the restriction of coral growth by the sediment, and in 1910 Wood-Jones similarly concluded not only that sedimentation was all important in atoll formation but also that it is "the real factor in the formation of barrier reefs." He believed that corals would colonize "any platform that rises to within the area above the *Limiting Line of Sedimentation*." Thus barrier-reef lagoons would be kept free of corals and colonized only on the outer margins.

This hypothesis could hardly have been erected with the Queensland Barrier Reef in mind, especially the northern sector, where there are fringing reefs, not everywhere inhibited by the mainland sediment; an inner channel, with a definite row of island reefs; a massive zone of inner reefs; an outer channel; and an outer barrier, which is not free from terrigenous sediment. If sediments inhibit coral growth to the extent envisaged by Wood-Jones, then why is the fringing reef almost continuous along parts of the northern sector? Why are the Inner and Outer channels practically open, but the middle zone almost blocked by enormous reef platforms? Why should the outer edge of the shelf be freer from sediment than other places and hence favor colonization by corals? On other continental shelves, considerable sediment is recorded passing over the shelf edge in many places.

Daly (1910, 1915, etc.) next postulated that, when the Pleistocene sea level dropped, the contemporary wave action "stirred up immense quantities of mud and sand that had been deposited during the previous period," and corals could not survive "such muddy asphyxiating baths." With mud control and lowered temperatures in the marginal latitudes, corals were essentially eliminated from most of our shelf areas. Beach rock, however, tended to form from the calcareous sand beaches, and on this firm foundation and clearer environment corals were able to grow up once more during the interglacial and postglacial rises of sea level.

There is some evidence in favor of this idea. The Great Barrier Reef borings showed an indurated littoral sand at the base of the coralliferous sediments. Muddy sediments are not found with the terrigenous basement sands. According to Daly, the fine materials would have been carried off the shelf into deep water. Admitting some degree of mud control in coral ecology, the criticism by Ladd and Hoffmeister (1936) stands "that Daly has greatly overemphasized the importance of such control." Serious problems which arise from Wood-Jones's theory are not explained any better by Daly's. In short, the factor of sedimentation, either Pleistocene or postglacial, does not explain satisfactorily either the distribu-

tion or the development of the Queensland Barrier Reefs.

One cannot ignore the role of Darwinian tectonic subsidence and glacial eustatic submergence in the origin of barrier reefs and atolls, although one must agree when Umbgrove (1947, p. 757) says that "sedimentation plays a very important part in the morphology of reefs." Were it not for the inhibiting effects of sediment in the shallow Pleistocene lagoons as postglacial submergence set in, corals would have colonized the areas behind the incipient barriers, growing upward from their beginnings as fringing reefs.

One may conclude that in the Queensland region, although the Recent shelf sediments are mainly terrigenous, they include locally numerous irregular lenses of organogenic reef sands and limestones. There is no solid mass of coralline material extending from the coast to the edge of the continental shelf, as has often been implied. The coral reefs of the Great Barrier complex could not have grown up from the floor of the lagoon under contemporary conditions of sedimentation and concomitant light restriction. The relative height of land and sea must have been reduced to permit the initial growths.

Interreef sedimentation now appears to be extremely rapid. Steers (1929, p. 243) noted that the smoothness of most of the lagoon floor points to active sedimentation. The failure to trace Pleistocene stream patterns from the main river mouths to the reef openings (Bryan, 1928) is also evidence in favor of rapid sedimentation. The deep borings at Michaelmas Cay and Heron Island show postglacial upgrowth of coral of 400–500 feet. Because the shelf beside these reefs is only about 15–25 fathoms (90–150 feet) in depth, some 300–400 feet of sediment must have accumulated on the shelf during the last hundred thousand years.

Normally, much sediment escapes from the shelf and is deposited on the continental slope. On the Queensland shelf, however, most of the sediment is trapped in the interreef areas. The great lagoon of the barrier reefs is filling up, just as are the small lagoons of the shelf atolls. The lagoon is shallower in the northern sector, which is narrower and has a much better barrier than the southern.

The Admiralty charts of the continental slope areas off the outer barriers show deep-sea deposits, especially off the northern sector, and not, as a rule, terrigenous deposits, such as might be swept off the continental shelf by currents. Close to the reefs, however, dead coral and coral sand may be found down to several hundred fathoms. Exceptional points are opposite the main openings in the reef, for example, Trinity Opening, and farther south, where there appear to be little deltas of muddy deposits deflecting the 200- and 300-fathom contours a mile or two farther out than elsewhere. The shallow parts of the broad "Coral Sea" platform (which extends at about 200 fathoms from Swain Reefs to Marion Reef and slopes away to the northwest) are covered mainly by clastic sediments, apparently terrigenous sands. These may be a residue after removal of all fine muds and oozes by the strong, southerly, longshore current.

Carpentaria, Arafura, Sahul, and North-West Shelf areas.—The northern shelf regions of Australia have fewer coral reefs than the Queensland shelf. The Gulf of Carpentaria has the least and the Sahul Shelf the most (Teichert and Fairbridge, 1948). Information on the sediments of the Sahul Shelf, and

adjacent areas, off northern Australia is meager (Fairbridge, 1950*d*).

While sedimentation is rapid in the Great Barrier Reef area, it is slow on the northern shelves. Few large rivers debouch onto them, and these drain a semi-arid hinterland, although the coast itself is tropical. There is no barrier on the outer edge of the shelf to retain the sediments, and the other reefs are too scattered.

Coarse "coral" sands occur close to the reefs and grade outward into finer sands and a green, sandy ooze on the shelf. The only detailed analyses are from three stations (188–190) of the "Challenger" Expedition, taken between Torres Straits and Aroe (Murray and Renard, 1891), which show typical "coral" sands near the reefs, whereas out on the open shelf there were quartz-glauconite sands with up to 38 per cent carbonate material, mainly molluskan debris and Foraminifera.

The quartz-glauconite sands are reminiscent of the terrigenous sands found in the Great Barrier Reef borings below 400 and 500 feet and suggest extremely slow or even suspended deposition over a very broad area. Most observers agree that glauconite formation is a slow process and that rapid burial inhibits it. Heavy residuals and only small amounts of deeply weathered feldspars suggest that these sediments have been deposited and reworked by waves and currents over a very long period, including periods of Pleistocene emergence.

Geological evidence (Fairbridge, 1950*d*) suggests only a relatively thin veneer of younger sediments on this shelf region, beneath which lie Mesozoic, Paleozoic, or even pre-Cambrian rocks.

In the Gulf of Carpentaria, on the other hand, much muddy sediment is carried down by the large rivers of tropical Queensland. This mud and other concomitant factors have largely inhibited coral growth here.

West-coast shelf area.—The western seaboard, south of North-West Cape, presents different problems. This is due to the arid nature of the coast and partly to the almost continuous lines of "sandstone" reefs. The Pleistocene calcareous eolianites of which they are made consist mainly of molluskan debris, Foraminifera, and generally about 10–20 per cent quartz and heavy minerals.

Erosion of the sandstone reefs provides most of the Recent sediments in this area. The balance is provided by invertebrate debris, Foraminifera, and, near the few river mouths, terrigenous material. Because the western part of the continent is largely an ancient and deeply weathered pre-Cambrian mass with its sedimentary derivatives, the principal minerals are quartz in varying stages of roundness, commonly wind-blasted, and the heavy residuals.

Exceptional areas of sedimentation include the Maud-Cloates lagoon (23° S.), where corals play an important role; Shark Bay (26° S.), where, in almost land-locked supersaline arms of the sea, $CaCO_3$ and probably other salts are chemically precipitated; Cockburn Sound, characterized again by calcareous muddy sediments and foraminiferal sands; Peel Inlet and other drowned river mouths, commonly closed by a sand bar for part of the year, where precipitated $CaCO_3$ and terrigenous deposits are found together.

The Houtman Abrolhos Islands on the outer part of the shelf, a coral group, are unique. In the Pelsart lagoon, coral sand and debris are constantly washed in over the windward reef. On the broad, flat reefs inside the lagoon, $CaCO_3$ has been

observed in the process of precipitation, apparently under the sharply rising temperature in the heat of the day. Banks of this white mud may be seen in protected bays and may also be found in the deeper spots of the lagoon (Fairbridge, 1948*a*).

On the continental slope outside the Abrolhos reefs, coral debris and sand reach down over 100 fathoms, but on most of the shelf there are fine to coarse quartz sands, in some places with dark residual minerals, grading into calcareous sands near the eolianite reefs.

HARD BASEMENT OF SHELF

It is assumed that beneath the younger sediments of the Queensland Shelf there is a hard basement of granite and metamorphic rocks of Paleozoic or greater age. It is exposed on the continental islands which lie on the shelf, such as the Lizard group, which is only 10 miles inside the outer barrier. Nevertheless, a thickness of not less than 700 feet of younger sediment is proved by boring at Heron Island, and elsewhere the floor is mainly covered by modern sediments.

The shores of both mainland and continental islands are cliffed and drowned, representing remnants of a once deeply eroded landscape. The base level of that ancient surface now lies well below sea level.

On the mainland, borings and wells in the younger coastal sediments disclose a rock floor at considerable depths. For over 400 miles the eastern coast is fringed by an alluvial flat, varying in breadth up to 30 miles or more. Borings have proved this flat underlain by alternating beds of clay, sands, and gravels, in many places reaching to over 100 feet below sea level (Jack and Etheridge, 1892). Their age has not been demonstrated by fossils.

Farther south, outside the barrier-reef area, at Brisbane, which is a moderate distance inland, borings showed the bed of an old river at exactly 100 feet below low-water springs (Richards, 1931). Still farther south, in New South Wales, in the region of Newcastle and Sydney, the rock floor has been traced down to a depth of about 250 feet below sea level (David and Halligan, 1908; Browne, 1945; L. A. Cotton, 1947). The harbors and river mouths all along this coast line are drowned by later submergence (Tenison-Woods, 1880).

This old erosion surface is generally correlated with one or a combination of the low eustatic sea levels of the Pleistocene, although this assumption may be open to grave doubts, as will appear below. The very late submergence of this coast line is supported, apart from coral-reef evidence, by the discovery of beach rock at 70–75 fathoms (420–450 feet) along the edge of the continental shelf of New South Wales (Smith and Iredale, 1924; L. A. Cotton, 1947), corresponding to similar coarse sediments on the Queensland shelf border (Bryan, 1928). This beach rock contains an abundant littoral fauna, corresponding to that of the present cold-water environment of Tasmania. The exposure of what is probably a late Pleistocene deposit on the shelf edge may be due to its being swept clear by currents and possibly also by gravitational slips of soft, younger sediment (Fairbridge, 1946). The latter would not be consolidated because it has not been exposed to subaerial or intertidal hardening.

There is little trace of Pleistocene drainage patterns or submarine canyons on this southern shelf. Hedley (1910) concluded that they would have been obliterated by the deposition of sediment caused by the powerful "Eastern Australia" or "Notonectian" current. Similar conclusions were reached in the Queens-

land area, although, in New South Wales, sediment collection by reason of a barrier reef is not possible.

Eustatic lowering corresponding to a 250-foot submergence in the Pleistocene is widely recognized. Smith and Iredale (1924), however, ascribe the beach rock at the edge of the New South Wales shelf to a eustatic low of over 400 feet. This would seem exceptional, and it may be related here, as in Queensland, to some tectonic depression. Wilkinson (1887) thought this abrupt shelf edge due to late Tertiary faulting, and Hedley (1910) took it to be involved in a broad asymmetric down-warp of the Tasman basin associated with faulting. (In his wide review of continental slopes, Gardiner [1915] included it among the typical fault scarps.) Recent faulting along such a steep edge, 7°–14° in places, is, according to Guppy (1890), unavoidable. This has been the opinion of apparently all other observers (David, 1911; Jensen, 1911; Bryan, 1944).

The problem of whether these low Pleistocene sea levels could explain the widespread buried erosion surface of the coast, that is, apart from local down-cutting of river channels and the beveling of soft, younger sediments, has been a subject of some debate. Daly (1910) made a major point of it, but later work has tended to discount the emphasis he placed on Pleistocene erosion (Ladd and Hoffmeister, 1936). The short duration of the periods of low sea level would not permit widespread peneplanation of hard rocks. Accelerated erosion of river gorges is all that seems possible in the available time.

The formation of this surface must, therefore, be older than the brief periods of Pleistocene low sea levels. It is probably related to the widespread peneplanation of eastern Australia, which culminated in Tertiary times (Andrews, 1902, 1910; Sussmilch, 1923, 1930). The low elevation of this surface must be due to widespread warping and fracturing during late Tertiary ("Kosciusko" phase of Andrews, 1910).[15]

Study of the physiography of eastern Queensland is essential to an understanding of foundations of the reefs. Jukes (1847) believed that an immensely thick wedge of coral actually formed the shelf. Darwin also believed that there was no true continental shelf beneath the barrier and that the floor of the lagoon had been filled up with thousands of feet of coral debris and associated sediments. But the continental shelf extends southward along the coast, even where there is no reef-capping. This concept also pays little heed to continental islands like the Lizard group. Rattray (1869) recognized the block-faulted nature of the adjacent coast and the probable relationship of the barrier reefs to the down-faulting of the underlying shelf.

Guppy (1890, p. 59) noted that there should be a great difference in contour between the northern part of the shelf, which has been reef-incrusted for unknown ages, and the southern part of the shelf, which is destitute of reefs. After examining the depths and widths of the shelf, he found no adequate distinction between the northern and southern sectors and concluded that the shelf extends along the entire eastern seaboard. Reefs merely form a capping on the northern sector. The reefs, therefore, would not be much thicker than the depth of the shelf.

Agassiz confirmed the observation that the continental shelf possessed the same general character both where it was reef-bearing and elsewhere. He rejected Darwin's and Jukes's idea that the entire bank was of coral. He pointed out that a subsidence of a mere 100 feet would adequately explain the major openings in the outer reefs opposite the big rivers of the mainland by fluvial erosion.[16] A subsidence of

[15] The interpretation of eastern Australian geomorphology by Sussmilch and others as due to block-faulting has lately been slightly modified; the fault blocks are still recognized for the most part, but some of the scarps are now considered to be fault-line scarps (C. A. Cotton, 1949).

[16] Hardly any sign of drowned valleys can be traced across the shelf (Bryan, 1928), as has been previously pointed out.

this order was indicated by the depths of the coastal-plain deposits and the channels of rivers at Townsville, Brisbane, and elsewhere along the coast (Agassiz, 1898, p. 111). On the other hand, he attributed the widespread evidence of the mid-Recent emergence to tectonic elevation and considered any idea of tectonic subsidence contradictory. Thus in any case he did not think that a barrier reef required subsidence to explain it. In fact, he believed that the continental shelf had been more or less stationary since the Cretaceous. He concluded that the reefs probably represent only a thin veneer on the shelf, some 20 fathoms (120 feet) in thickness (p. 139).

Considerable opposition to Agassiz's views was expressed at the time, not all of it justified. For example, Daneš (1911) stated that Agassiz was flatly opposed to subsidence, which was true as far as several thousand feet are concerned but was still not strictly correct. Andrews (1902), after studying the physiography of the adjacent mainland, concurred essentially with Agassiz in believing that the reef was growing from a pre-formed platform. He believed that it grew up from an old shore line, as shown by its parallelism with the present coast. The Guppy-Agassiz view, that the shelf has the same form whether or not corals are present and proving that corals are essentially a veneer, was confirmed. Andrews showed a section of this platform veneered by not more than 300 feet of late Tertiary sediments, surmounted by a series of scattered reefs not more than a few hundred feet thick, reaching a maximum of 600 feet in the outer barrier. Even with more recent data, it would be difficult to improve on this general concept.

Hedley and Taylor (1908) admitted the possibility of a ready-prepared platform, but at the same time they insisted that subsidence of some 200 feet was required to explain the buried channels and the drowned coast line and islands: in short, a modified Darwinian viewpoint. To this Daneš (1911, 1912) added evidence on the extensive peneplain of central Queensland, which extends out under the continental shelf, and its complex displacement by youthful block-faulting. Jensen (1911) illustrated the latter as one of a number of antithetic fault blocks, stepping down to the east.

Davis (1917) concurred with Darwin and traced the history of the reef back through many subsidences to the Cretaceous. Davis' extension of this Darwinian subsidence theory on a world-wide basis in his "The Coral Reef Problem" (1928) is the rather one-sided product of this *idée fixe*. Davis failed to take full account of glacio-eustatic oscillations of sea level, the principles of which had been demonstrated by Tylor, Penck, and others. Daly's theory of glacial control (1910, 1915, 1925, 1929, 1934, 1948, etc.) was seized upon by Andrews (1922), who accepted the idea of some Pleistocene erosion in the formation of the shelf, so that the upgrowth of the reef was mainly postglacial, instead of older as he had previously thought. He pointed out that, whereas Tertiary coral limestones were generally well-bedded outwash deposits, with very little massive coral *in situ*, the Pleistocene (and Recent) coral limestones were not bedded, having grown rapidly upward during the recent submergence.

Vaughan (1919, 1923*b*), following Andrews' idea of the reef foundation, revived the Wharton concept of the antecedent platform which could be lowered relative to sea level either by tectonic subsidence (modified Darwin hypothesis) or by eustatic rise of the ocean (Daly hypothesis). Vaughan, like Agassiz and Andrews, noted that the shelf profile was materially the same whether corals were present or not and that in many places the barrier reefs did not rise from the outer margin of the shelf but were back some distance. He also pointed out that, where there are gaps in the outer barrier, one might expect deep re-entrants if the shelf were entirely of coralline origin. Absence of such re-entrants denies this concept.

Hoffmeister and Ladd (1936, 1944) favored the same antecedent platform but believe that, ecological conditions permitting, a reef will build up from any pre-formed platform *without* change of sea level. Theoretically, this may be so, but there has been a world-wide 300-foot postglacial rise in sea level, and ecologic conditions in the Barrier Reef region do not permit corals to grow up from more than 6–7 fathoms (36–42 feet). Because shallow-water reef corals are found in the borings at 400 and 500 feet, a certain amount of tectonic subsidence is also unavoidable.

Thus we come back essentially to the views of Andrews and Vaughan. These ideas were substantiated by the physiographic studies of Stanley (1928*a*, *b*), who also concluded that the present coast line and continental-shelf base-

ment were initiated by fracture and subsidence in Tertiary times. During the Pleistocene, oscillations of sea level occurred, but reefs continued to grow on the shelf, probably as far south as Cairns, and a considerable accumulation of sediments took place on the shelf in spite of periodic emergence and erosion. This aggradation still continues rapidly today. The bulk of the surface reefs grew up within the postglacial era. Physiographic work by Richards, Hedley, Jardine, Steers, Sussmilch, and others has amplified the above-outlined scheme. The history of the Great Barrier Reef, at least of the southern half, is thus gradually becoming clearer.

There is no longer, really, an over-all "barrier-reef problem." Its structure and origin are now pretty well understood. Although there are many associated problems and questions upon which more data are desirable, the mystery is no longer an all-embracing one.

On the other hand, the character of the *basement* of the Queensland Shelf and Coral Sea is still unknown. Hypotheses based on topographic and geologic evidence are unproved because neither of the Great Barrier Reef borings reached the basement and the continental islands do not lie far from the mainland.

As pointed out elsewhere, the topographic character of the Queensland Shelf and its adjacent borders in the Coral Sea led to the conclusion that tectonic subsidence with block-faulting initiated the present shelf in late Tertiary times. Further subsidence of the shelf occurred as recently as late Pleistocene. The steep character of the outer edge of the shelf requires major faulting along this line, but, as seen on the map, this steep slope dies out gradually to the southeast and is scarcely recognizable at the latitude of Swain Reefs. The floor of the Coral Sea is thus separated from the Queensland Shelf by a gigantic hinge fault, sloping down to the north and gradually increasing from a few hundred feet in the southeast to some 6,000 feet in the northern sector of the Great Barrier Reef.

From the shallow end of the hinge, the sea floor drops away very gently to merge with a great ovoid bank from which rise the Saumarez, Flinders, and Osprey atolls of the Coral Sea. This bank, which is nearly 1,000 miles long and up to 300 miles wide, we propose to name the "Coral Sea Platform." Its surface is almost flat, but it slopes down very gently from south to north.

Except at the hinge end from Swain Reefs to Saumarez, it is separated from the Queensland Shelf by a long, narrow depression about 50 miles wide, which we suggest be called the "Queensland Trench."[17] It seems to be cut off from the Coral Sea Platform by a normal fault system parallel to that marking the edge of the Queensland Shelf, thus corresponding to a deep, submerged graben.

Beyond the Coral Sea Platform there is a broad trough parallel not to the Queensland Shelf but to the young mobile belt of New Guinea–New Hebrides. It seems to be an extension of what is known as the Papuan Geosyncline of the southern side of the New Guinea "backbone" and is accordingly named the "Papuan Trough" (fig. 10).

Evidence for regional tectonic subsidence of the Coral Sea Platform, in-

[17] The writer follows the recommended geographical method of oceanographic nomenclature favored by Vaughan (1940) and others; also the system of rank terminology, except that "platform" is preferred to "plateau." In this way the "Coral Basin" of Supan, Wüst, and others is subdivided on both geologic and bathymetric grounds into three: "Papuan *Trough*," "Coral Sea *Platform*," and "Queensland *Trench*."

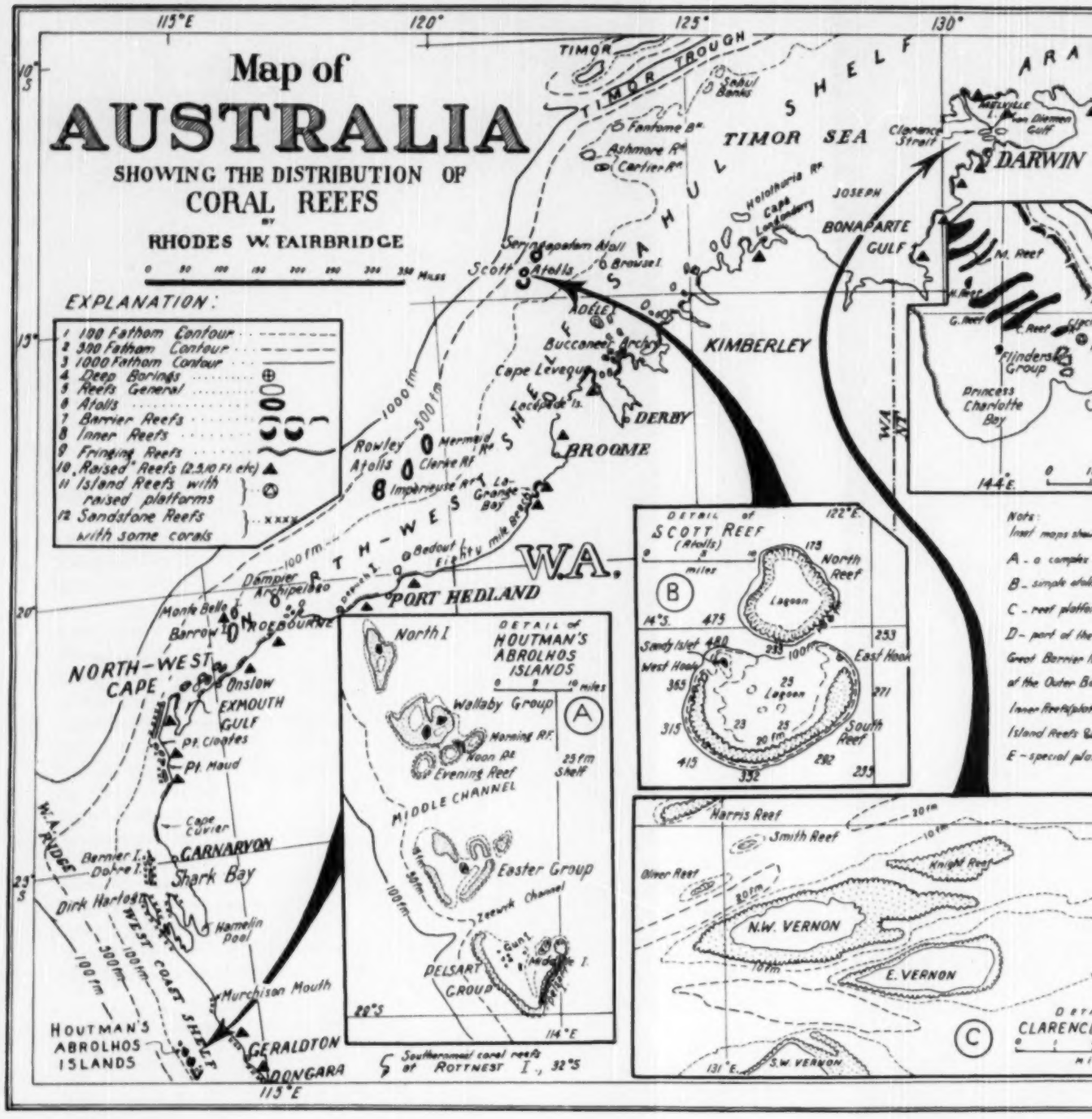
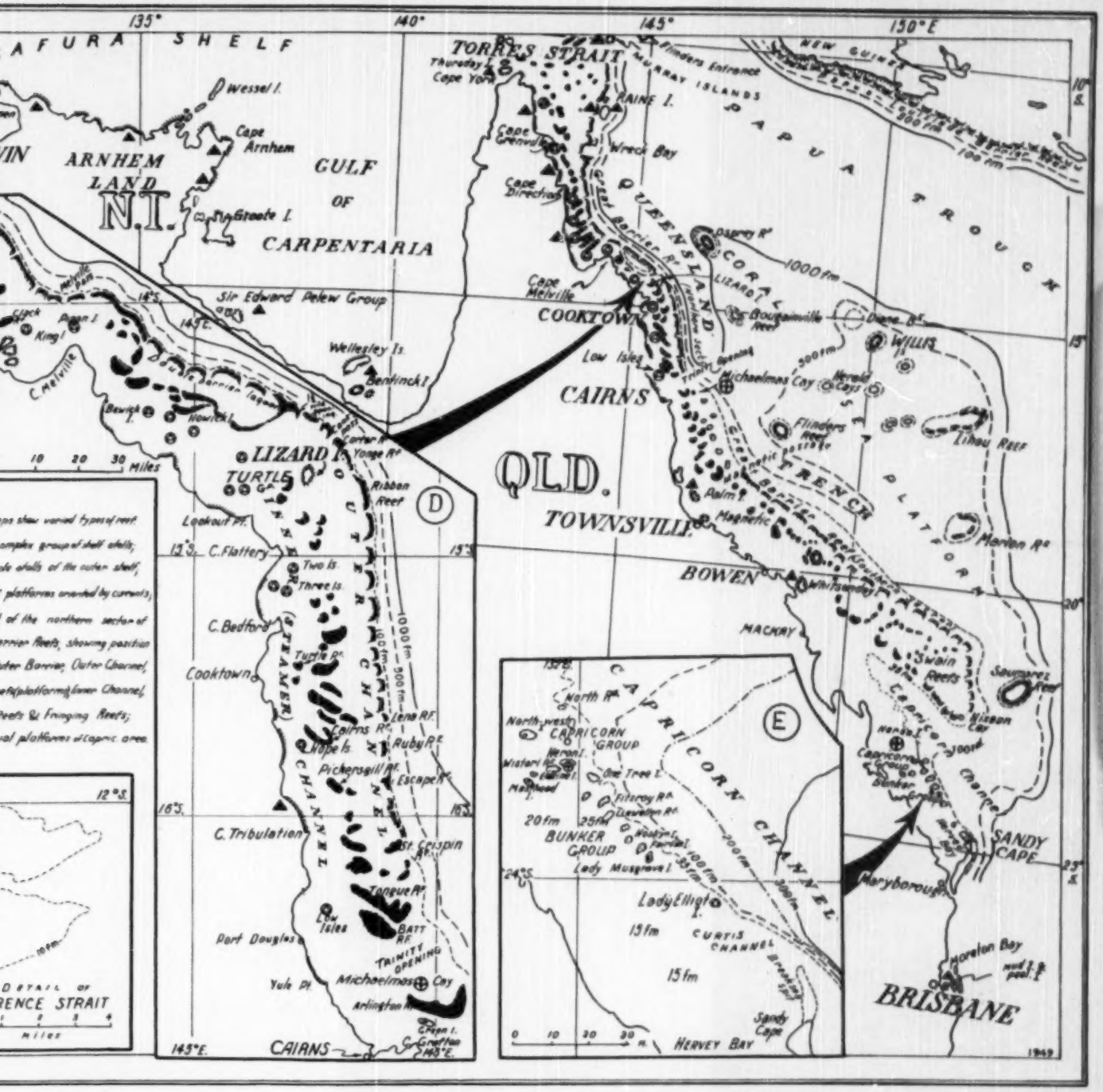


FIG. 10.—Map of Australia, showing the distribution of coral reefs, including insets of the Houtman Abrolhos



olhos Islands, Scott Reef, Clarence Strait, Great Barrier Reef (northern sector), and the Capricorn-Bunker group

volving a northward tilt, in a trend almost parallel to the main coastal trend is hardly deniable. Whether the cores of the atolls arising from it are volcanic or have grown up from the reef patches during subsidence is uncertain. The position of the principal reefs, however, near the margins of the platform suggests submerged fringing reefs. The small reefs rising from the central parts of the platform are concentrated near the 400–500-fathom lines, which cross the platform as it is tilted northward and might, therefore, be a former shore line.

Paleogeographic data suggest that the subsided area is a part of the former continent of Tasmantis.[18] Bryan (1928) believes that the 100-fathom contour on the edge of the shelf north of Swain Reefs is a structural line of great antiquity and importance, being the former border of the Tasman Geosyncline of eastern Queensland and the ancient continent of Tasmantis to the east, whose dismemberment began in early Tertiary times (Bryan, 1944).

Bryan pointed out (1928) that the central axis of the Tasman Geosyncline of Paleozoic times in eastern Queensland lay more or less along the trend of the present coast ranges; its western borders lay in the vicinity of the present continental divide, whereas the eastern part is missing. The western half of the geosyncline is narrowest in the north and broadens out gradually toward the south. The shape of the continental shelf is almost a mirror image of this pattern and may be the "missing" half. One may add that the big inward swing of the shelf at Sandy Cape is paralleled on the west by a great embayment of younger Mesozoic rocks on a cross-axis approximately through Brisbane. Bryan concluded that, because the highlands on the west of the axis consisted of Silurian-Permian of north-south trend, one might expect to find similar rocks beneath Swain Reefs but in northwest-southeast trends.

The deep, gulflike Capricorn Channel occupies a position in the south which is almost matched by the Queensland Trench in the north (fig. 10). It is marked by the 35-fathom contour swinging in more than 100 miles onto the shelf from the southeast; its axis lies in the northwest trend of the adjacent mainland structures. It is, however, a more gently contoured trough than the steep graben of the Queensland Trench. Bryan (1928) suggests that it may be an extension of the Maryborough synclinal basin of Mesozoic rocks. However, it does not lie in a direct line and may, therefore, be an *en échelon* structure related to the Maryborough depression. Bryan pointed out that the repeated cable breaks in the extension of this trough, north of Breaksea Spit, might be due to faults and slumps in the Mesozoic basement. Earthquakes and displacements of the 100-fathom line have been reported over some thirty-five years, but restudy shows that many data concerning them were inaccurate (Bryan, 1936). However, the submarine contours between the Bunker group and Sandy Cape come very close together, so that it is logical to assume a fault scarp here. Off Sandy Cape itself there is a drop of over 6,000 feet in a very few miles. If there is a fault, it is hinged down to the southeast from the vicinity of the Capricorn-Bunker groups.

The 1,000-fathom contour, after skirting the Saumarez Reefs, swings to the southeast close to the 300-fathom line,

[18] The eastern border of Australia has been considered as a down-faulted and subsided subcontinent, to which the name "Tasmantis" was given by Sussmilch and David (1919). Further discussions on this subject may be found in Marshall (1910, 1933), Jensen (1911), Gregory (1912, 1930), Sussmilch (1938), and Bryan (1944).

which outlines a submerged ridge extending 110 miles southeast of Swain Reefs before it doubles back 20 miles in the direction of the Capricorn Channel and eventually turns south to parallel the coast past Sandy Cape. Similarly, all the shallower contours curve back into the "Capricorn Trough," as we may call this great synclinal structure, which is thus plunging to the southeast.

In this submerged trough, the 100-fathom line is of no particular significance; the floor of the depression slopes down steadily to the southeast from 21°20′ S. without any significant structural break to the 2,000-fathom line, 270 miles from the northwestern end. The 2,000-fathom contour delineates a great area of subsidence of the northern Tasman Sea leading directly into the northwesterly trend of the Capricorn Trough. Structurally, it is impossible to locate the true edge of the present continent in this area. The slope of the Capricorn Trough descends gently down to part of the subsided ancient continental mass of Tasmantis. The continental shelf is cut off fairly sharply along most of the outer barrier, and also south of the Capricorn Trough off Sandy Cape. In such areas we agree with Bryan (1928) that "the 100 foot contour thus represents the real structural limit of the continental mass . . ."; but, as Spender (1930, p. 279) pointed out, this seems not to be true in the Swain Reef–Capricorn area, where the steep slopes are down somewhere near the 1,000-fathom mark. Soundings, however, are sparse.

An analogous situation exists in the Sahul Shelf region, where maps depict the 100-fathom line as the continental border, whereas atolls and reef platforms rise from near the true edge at about 300 fathoms. Subsidence along this tract, facing the mobile East Indian arcs, is suspected (Teichert and Fairbridge, 1948).

The question of the shelf floor and its hard basement off the northern and northwestern coasts, briefly considered below, is treated separately in detail (Fairbridge, 1950*d*).

From the character of coral growths and from the tectonic evidence, the continental shelf appears to be subsiding fairly rapidly in two distinct areas, whereas in one large area it appears to be rather stable. The latter, the Shark Bay–Abrolhos–Rottnest sector, or "West Coast Shelf," does not exhibit Recent atolls, barrier reefs, or a deeply depressed outer margin. On the other hand, the North-West–Sahul–Arafura shelf area appears to be subsiding locally, corresponding to the paralic basins, whereas the Queensland Shelf is subsiding marginally in a more uniform way, although apparently more in the south than in the north.

There has been considerable discussion as to the nature of such movements of the continental shelf areas. The peculiar topography of certain shelf areas can be explained only by subaerial forces in combination with shore-line erosion (Fairbridge, 1950*b*), and their present submergence is due either to complex eustatic causes (Bourcart, 1938; Shepard, 1948) or to tectonic down-warping (Francis-Bœuf, 1938; Jessen, 1943; Umbgrove, 1946, and others).

Eustatic oscillation would allow an emergence of the continental margins of only some 100 meters or so (Umbgrove, 1930*a;* Daly, 1934), and this would be inadequate to explain the deep foundations of the Australian shelf reefs (some over 1,500 feet), not to mention the deeper mid-oceanic ones (Molengraaff, 1917; Stearns, 1946). The distribution and depth of shelf reefs suggests that it is

a marginal down-warp and not a giant eustatic oscillation, such as visualized by Shepard (1948). Tectonic evidence almost all around the Australian coast line suggests recent normal faulting.

THE BARRIER-REEF BORES

Summary of results.—The first Australian reef boring was put down at Michaelmas Cay under the auspices of the Great Barrier Reef Committee in 1926. This reef is an intermediate platform, lying 22 miles off the mainland near Cairns and 10 miles within the 100-fathom line at 16°36′ S. (see figs. 11, 12). The cores disclosed 404 feet of soft, unconsolidated coral and coralline algal debris without any dolomitization or extensive hardening but with thick intercalations of foraminiferal and coral sand. Below this was a rounded quartz-glauconite to calcareous sand, grayish-green in color, which contained numerous Foraminifera and a small amount of pyrite; the boring was abandoned in this material at 600 feet. Between 404 and 476 feet there were also coral fragments, indicating reefs in the vicinity. Most of the analyses down to 404 feet showed over 95 per cent carbonates, whereas in the lower part the carbonates drop sharply to less than 50 per cent (Richards, 1928; Chapman, 1931; Richards and Hill, 1942) (fig. 11).

A second borehole put down in 1937 was located some 700 miles southeast of the first, at Heron Island (23°26′ S.), some 44 miles from the mainland and 10 miles inside the 100-fathom line. The cores disclosed a similar section (see figs. 11, 12). From 0 to 506 feet there were coral-reef and foraminiferal sand facies. This graded abruptly into quartz-foraminiferal sands from 506 to 732 feet, intercalated here and there by "hard bands," probably travertinized calcareous sand or beach rock. The Foraminifera were generally infilled by glauconite (Richards, 1938*a*, *b*, 1939*b*; Cushman, 1942; Iredale, 1942; Richards and Hill, 1942).

The core recovery and sampling was much more complete in this second hole, so that the two logs are not precisely comparable. Major correlations are apparent. Discussion of the more detailed features will follow below.

Comparisons with other bores.—Richards showed that the logs were similar to those of other shelf-reef areas, such as the Florida shelf (Sanford, 1909; Vaughan, 1910; Matson and Sanford, 1913; Cooke, 1945). In the Key Vaca boring, there are 105 feet of the white Key Largo limestone, partly coral rock and partly a lagoon mudstone of the rock type commonly found in the Queensland borings and also characteristic of the Abrolhos. At Key Vaca, Cooke believes that the formation belongs to the third interglacial period. This Key Largo limestone rests on shelly marl and limestone, which may be earlier Pleistocene. The underlying quartz sands, with some glauconite and clay bands, appear at 176 feet and have been followed to 700 feet. Their age may be Pliocene or Miocene. A rather similar section was disclosed in several other borings on the Florida Keys, although at the southwestern extremity, Key West, there was much less quartz sand and more marl, limestone, and calcareous sand, down to 2,000 feet (Hovey, 1896; Matson and Sanford, 1913).

Comparisons with borings made in a noncontinental-shelf area, such as Hawaii, are likely to be less significant, but even here a fairly high degree of stability seems to have existed during and since the Pleistocene, so that the almost continuous logs of "coral" in many of the bores above 200–300 feet are suggestive

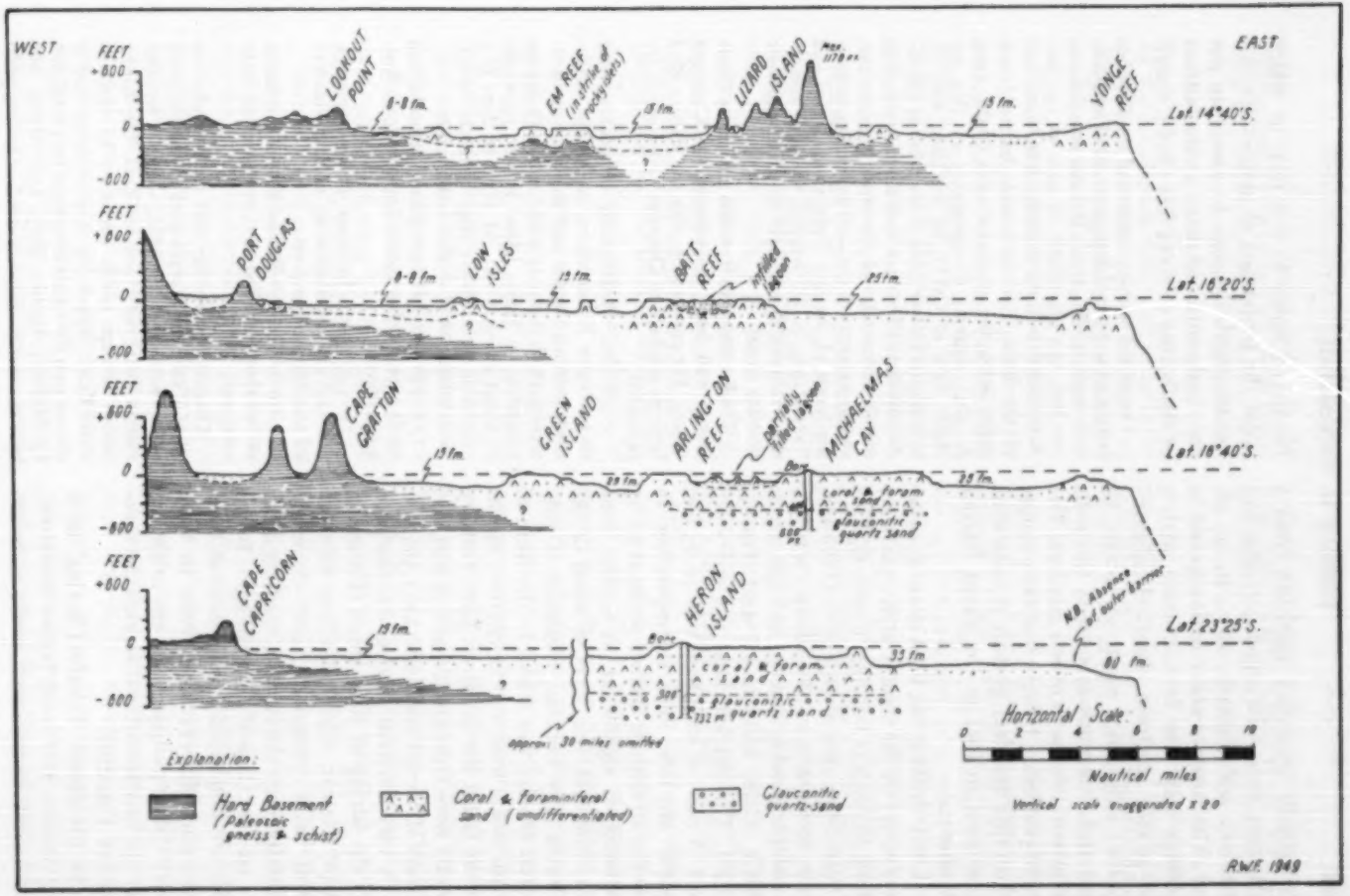


FIG. 11.—Sections of the Great Barrier Reef, showing relationship of continental islands and known information about the basement, as shown by bore logs at Michaelmas Cay and Heron Island.

of uniform (eustatic) upgrowth from a platform now submerged 300–360 feet (Stearns and Vaksvik, 1935, p. 39, pl. 20). The bases of other reefs exposed in borings may also be correlated with a 60–90-foot platform observed offshore. Earlier subsidence must, of course, be assumed, inasmuch as some of the wells show coral down to over 1,000 feet. The correspondence, however, of the younger platforms and reef growth is somewhat analogous to that in the Great Barrier Reef area.

Little drilling has been done in other reef areas. Of the oceanic atolls, there are three significant borings: Funafuti, Kita Daito Jima, and Bikini. Unfortunately, only summaries of the Bikini cores are available (Ladd, Tracey, and Lill, 1948; Wells, Cole, Lill, Ladd, and Tracey, 1948). The deepest hole reached 2,556 feet. A zone of hard bands at 35–65 feet below low-tide line, corresponding in part to an offshore terrace of about 10–15 fathoms, is significant. A similar zone was found at this level in some of the Florida Key borings. Evidence of subaerial erosion was found in the form of fossil molds from 170 to 575 feet, just as in the Great Barrier Reef. Late Tertiary corals were encountered at 930 feet and Lower Miocene Foraminifera at 1,305 feet.

A very different section was disclosed by the boring at Kita Daito (Hanzawa, 1940; Nugent, 1948). This is an elevated atoll, with phosphatic and dolomitic limestones on the surface, reaching down to 340 feet (Plio-Pleistocene age). Below this are noncoralline sediments of Miocene and Oligocene age, down to 1,420 feet. This is perhaps what one might expect to be encountered *beneath* the dolomite of Funafuti.

In the classic Funafuti boring, as is well known, reef coral, lagoon limestone, coral, and foraminiferal sands were encountered down to 638 feet, at which point the proportion of magnesium carbonate suddenly rose to about 40 per cent and continued down to the bottom of the boring at 1,114 feet (Judd, 1904).

One of the striking features of the Funafuti boring is the gradual disappearance of aragonite with depth (Cullis, 1904). It is common down to 100 feet, but practically all gone at 150 feet. None occurs with the dolomite in the lower half of the hole. The cavernous, leached cores found below the aragonite levels (Judd, 1904, p. 384) suggests disappearance of this unstable form of $CaCO_3$ by leaching, with reprecipitation lower down in the form of calcite. According to the writer, such leaching would be mainly subaerial and therefore suggests successive exposure of the atoll during low eustatic sea levels of the late Pleistocene. David and Sweet (1904, p. 88) reached the same conclusions concerning sea level on geomorphologic grounds alone.

The Foraminifera were first studied by Hinde (1904) and later re-examined by Chapman (1944). For comparison with the Barrier Reef borings, points from Chapman's conclusions are presented.

0–550 feet.—Alternations of shallow- and deeper-water Foraminifera, suggesting oscillations between deeper and shallower environments, which compare with Cushman's similar comments on Heron Island boring, especially at about 450–500 feet. At 375 feet there is a suggestion of an eolian stage from the nature of accumulation of shallow-water forms.

550–750 feet.—Deeper phase, foraminiferal sands with only few corals. Temporary shallowing at 660 feet.

750–1,050 feet.—Evidence of slow accumulation and gradual shallowing at first, followed by a sudden submergence.

1,050–1,114 feet.—Shallow littoral conditions at bottom, followed by evidence of submergence.

Chapman points out that the reproductive stage of the large *Marginopora vertebralis*, found at the bottom of the hole, normally inhabits shallow coral pools and reef-flats. The boring passed through corals encrusted by shallow-water forms, supporting other evidence that it did not pass through a talus bank (as postulated by Murray, Agassiz, Daly, and others), but that the reef buildup occurred in a centripetal

manner, as believed by Darwin and Dana and demonstrated by Skeats (1918*a*). Hinde (1904) noted that shallow-water corals were in position of growth at the bottom of the hole. Also *Halimeda*, which normally live from 30 to 270 feet, were found all the way down to the bottom. All the Foraminifera were types living in the area today, but with varying abundance at different levels. Chapman believes that they could not be older than early Pleistocene.

The dolomite problem.—The findings at Funafuti focused attention on the problem of dolomite formation. Skeats made analyses of dolomites from elevated reefs in the Pacific and Indian oceans (1903) and investigated the Triassic dolomitic reefs of the Tyrol (1905). He came to the conclusion that marine dolomitization took place in shallow water of 0–150-foot depth, under slightly increased pressure, with abundant CO_2, and in a porous limestone (to allow free access of Mg-bearing sea water). The process occurred best under conditions of slow subsidence, giving time for the reaction to complete itself. He restated his case clearly in 1918 (*b*).

Since then further work has been done on the dolomite problem (summarized by Van Tuyl, 1916; Twenhofel, 1932). Most valuable support for Skeats's interpretation came from Udluft (1929), who confirmed the shallow-water environment and demonstrated that reducing conditions were indicated by bitumens and ferric iron, thus high CO_2 pressure.

Marshall (1930) carried out field tests on the *makatea*, or raised coral rim, of Atiu in the south Pacific, showing a "definite and gradual decrease in dolomitization from the inner toward the outer side of the makatea." Udluft's observations suggest that favorable reducing conditions might have existed within a lagoon such as that of Atiu when it was a normal atoll or barrier-reefed island. Such reducing conditions are encountered in the stagnant bottoms, rich in H_2S, of closed atolls today.

Clarke and Wheeler (1922) demonstrated that primary $MgCO_3$ is present in various marine invertebrates in very appraisable quantities. Thus many Foraminifera run 8–12 per cent $MgCO_3$, Alcyonaria 6–15 per cent, Bryozoa 2–11 per cent, whereas in calcareous algae, such as *Lithothamnion*, it ranges from 10 to 25 per cent. On the other hand, madroporarian corals and Mollusca average only about 0.5 per cent. A typical "coral" sand containing one-quarter each of Foraminifera, madroporarian coral, calcareous algae, and molluskan debris thus may consist of about 50 per cent of material with an average of 10 per cent $MgCO_3$, diluted by 50 per cent with very little magnesia. The result is a "coral" sand of 5 per cent or less $MgCO_3$. Coral-reef rocks of this concentration are never dolomitized; as a rule, dolomite crystals start to appear only if $MgCO_3$ is over 15 per cent, pure dolomite being about 45 per cent $MgCO_3$. Secondary concentration would thus be necessary to transform the normal reef sediment into dolomite. Clarke and Wheeler suggested differential leaching under subaerial conditions. Most reef borings show evidence of this to some extent, but the removal of the tremendous bulk of $CaCO_3$ so required precludes it as a final explanation of reef dolomite.

There are, therefore, three recognized sources of $MgCO_3$-rich sediments in coral-reef rocks: primary organogenic secretion in shallow water; secondary inorganic enrichment from sea water, under reducing conditions below sea level; and secondary concentration under subaerial conditions. To these must be added the intertidal primary precipitation of $MgCO_3$ from sea water on the reef-flat under the specialized conditions of raised temperatures and pH (aided by photosynthetic production of CO_2 [Rivière, 1940]), together with secondary enrichment in the same zone by alternate solution and reprecipitation.

Returning to the Funafuti data, a critical re-examination of the $MgCO_3$ analyses was undertaken by Reuling (1934). He noticed an astonishing fact, namely, that the slight variations of $MgCO_3$ (about 3–5 per cent) in the upper part of the section (above 638 feet) are paralleled by variations in the lower half of the

boring, each separated from the other by a distance of 638 feet. He concludes, therefore, that dolomitization (over 40 per cent $MgCO_3$) is taking place today in a zone from a depth of 638 feet to about 20 feet lower.

In order to explain the massive column of dolomite from 638 down to 1,114 feet, he assumed progressive subsidence, just as Skeats had done in the Tyrol dolomites. Again, as Skeats had concluded, he believed that slow subsidence enabled the reaction to function nearly to completion (40–43 per cent $MgCO_3$), whereas acceleration in subsidence quickly sealed off the porous reef rock and caused the percentage to drop (in places as low as 5–25 per cent).

Reuling realized that $MgCO_3$ simultaneously accumulates in the intertidal belt, where the same rule will apply (stability: $MgCO_3$ over 5 per cent; rapid subsidence: $MgCO_3$ less than 5 per cent). Accordingly, stability will lead to high $MgCO_3$ figures both at the tide level and at −638 feet (say 10 per cent and 40 per cent, respectively). On the other hand, rapid subsidence will lead to low figures at each level (say 1 per cent and 25 per cent, respectively). When these variations are shown on graphs of an enlarged scale, the coincidences are very striking. For example, Reuling found that at −860 feet in the Funafuti boring there is only 20 per cent $MgCO_3$; at 860 less 638, thus at 222 feet, there is correspondingly only 1 per cent $MgCO_3$ in the contemporary intertidal accumulation.

Reuling believed that the particular figure of 638 feet at Funafuti was a function of the latitude (related to temperature and pressure). Thus, theoretically, the level of the dolomitization zone should rise *away* from the equator. The only borings, it seems, with $MgCO_3$ figures available are those of Key West and the Great Barrier Reef. All are farther away from the equator, yet, strangely enough, none of them shows any dolomitization. Key West shows a peak of 14 per cent $MgCO_3$ at 775 feet and Heron Island 16 per cent $MgCO_3$ at 700 feet, but these are probably only long intertidal stillstands. However, neither of these is an atoll.

Reuling's theory has not yet been tested on any other atoll borehole. Funafuti is a perfect atoll; Key West, Michaelmas Cay, and Heron Island are not and have never provided the stagnant closed basin that probably existed in the center of the Funafuti lagoon over a long period of its subsidence. Here perhaps would be the explanation for the absence of dolomitization in the Great Barrier Reef borings, which puzzled Richards and Hill (1942).

For the main Bikini boring, the brief abstract reports indicate no dolomitization (Ladd, Tracey, and Lill, 1948; Wells, Cole, Lill, Ladd, and Tracey, 1948). In this case we have a true oceanic atoll, which should theoretically correspond to Funafuti in a very close manner. The borings on the lagoon island of Maratoea, off Borneo (Kuenen, 1947), also lack complete reports, but there is no mention of dolomite as yet. If in both cases no dolomitization is proved, one may suggest that here were atolls (one oceanic, one shelf example) that may not have had the necessary reducing bottom conditions.

Discussion of Great Barrier Reef borings.—Both borings (fig. 12) penetrate the glauconitic sands at depth. These are almost certainly littoral and neritic sands of the Pleistocene shelf. Neither the Mollusca (Iredale, 1942) nor the Foraminifera (Chapman, 1931; Cushman, 1942) indicate a greater age.

In the Heron Island boring, hard

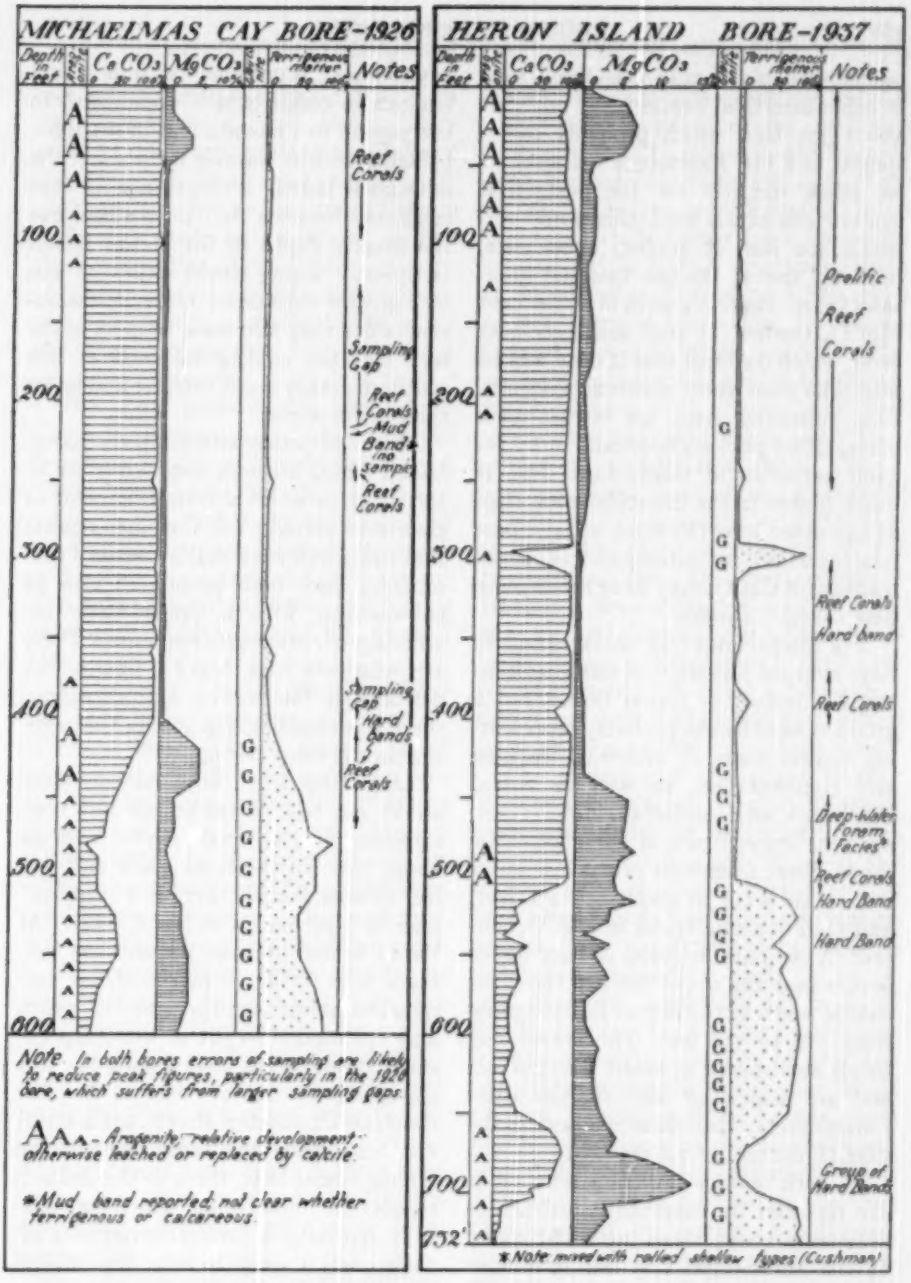


FIG. 12.—Comparison of Michaelmas Cay and Heron Island bore logs and analyses (based on Richards and Hill, 1942, etc.). Magnesium carbonate figures, as a sensitive indicator of intertidal deposition, are shown on ten times the scale of the calcium carbonate and terrigenous percentages. Note approximate correlations between 400 feet in Michaelmas Cay hole and about 500 feet in Heron Island bore, suggesting 100 feet greater subsidence in latter.

bands, probably of calcareous beach rock or other intertidal concentrate, occur at about 700 feet, which probably corresponds to a low Pleistocene sea level. If we allow 300 feet for the postglacial eustatic rise of sea level, this would still entail 400 feet of gradual subsidence, probably during the last hundred thousand years. There is a peak of 16 per cent $MgCO_3$ content at that same 700-foot level, which confirms that it once was an intertidal zone where chemical precipitation, primarily from sea water, took place, either physicochemically or by organic secretion of magnesium. This is much higher in the intertidal zone than at any other level (Reuling, 1934). Some concentration by subsequent subaerial leaching of $CaCO_3$ may have taken place here (Skeats, 1918*b*).

The Michaelmas Cay boring seems to have stopped just short of this lowest intertidal horizon of Heron Island, but a group of hard bands, probably representing similar zones of intertidal leaching and concentration, as well as actual beach rock, are found at 421 and 477 feet, near the upper limits of the glauconitic sands. Some fragments of shallow-water reef corals occur at about 406–476 feet, and there is a small peak in $MgCO_3$ content. A comparable group of hard bands occurs near the upper limit of the glauconitic sands in the Heron Island boring from 510 to 558 feet. The lowest reef corals also appear at about 506–526 feet and are associated with shallow-water Foraminifera. Also, there is a peak in the $MgCO_3$ content of 7.8 per cent.

In both borings the glauconite sands are replaced by dominantly calcareous sediments above these intertidal zones. In the Michaelmas Cay boring it is at 400–450 feet, in the Heron Island boring at 450–506 feet. In both cases a littoral deposit of terrigenous sands, a cold-water environment, is gradually swamped by a sea rich in corals, which would seem to correspond to a eustatic rise in sea level, bringing with it warmer waters and the associated fauna. A correlation is thus suggested between the two borings, but the greater depth in the Heron Island (southern) boring would imply 50–100 feet greater subsidence there, a conclusion which may find some support in the fact that the continental shelf in this southern area is much broader and lower than in the north.

Above the terrigenous sands the Heron Island boring is a zone approximately 50 feet thick in which a certain amount of glauconite persists, and Cushman reports that rolled and worn shallow-water Foraminifera have been introduced into an environment with a deeper-water assemblage of undamaged specimens. There are relatively high $MgCO_3$ figures (6.7 per cent) in this section. Unfortunately, there is a sampling gap just in this position in the other boring.

Above this level, both borings show about 100 feet of calcareous but noncoralline foraminiferal sands, and at about 250 (Michaelmas Cay) and 350 feet (Heron Island) there is a reappearance of hard bands and reef corals. At Heron Island, between 308 and 289 feet, there is a sudden invasion of fine terrigenous sands, mainly quartz, which may correspond to one of the sharp eustatic drops in sea level during the last glaciation. At 211–213 feet in the Michaelmas Cay boring, that is, again about 100 feet above the same level in the Heron Island hole, there is the driller's report of a "mud band," which may possibly represent a similar terrigenous invasion, but it seems to have been missed in the sampling.

From this horizon, up to 17 feet in the Heron Island cores are mainly massive

reef corals, partly cemented with lime mud ranging to a hard, creamy-white reef limestone, soft, partly cemented reef detritus, soft, limey marl, and "coral" sand. From 17 feet to the surface there is calcareous sand. In the upper 50 feet there is a peak of $MgCO_3$ at 6.7 per cent, suggesting a concentration by precipitation and leaching at sea level during the late Würm stillstands of the sea at about 10 and 3–5 fathoms.

The upper parts of the Michaelmas Cay hole are very similar, with reef coral limestone, loosely cemented debris, and "coral" sand. A similar $MgCO_3$ peak of 3.2 per cent occurs in the top 50 feet. In the Funafuti boring, likewise, there are peaks with 16–22 per cent in this zone. And 8–27 per cent was found in a core of beach rock at Bikini (Ladd, Tracey, and Lill, 1948).

In the top 100 feet of the Great Barrier Reef holes, aragonite is preserved, whereas, lower down, almost all is leached out or converted to calcite, presumably by subaerial circulating waters during some of the brief eustatic drops of sea level of the last glaciation. This tallies with the observations at Funafuti. There is a difference, however, in that aragonite is found preserved again in both borings in the lower sections, below 400 feet in Michaelmas Cay and 500 feet in the Heron Island hole. These would presumably have been located lower than any eustatic drop in sea level and consequently would have been spared any subaerial solution and diagenetic effects. At Funafuti the deeper parts of the borings may be younger than on the Great Barrier Reef, and subsidence may have been more rapid.

The negligible amount of dolomitization in both the Michaelmas Cay and the Heron Island borings is the most striking feature and contrasts with the Funafuti boring. It is perhaps explicable because closed lagoons under reducing conditions and increased CO_2 pressure, which would have favored dolomitization, did not exist here, as in the case of Funafuti (Reuling, 1934). On the other hand, the northern half of the Barrier Reef area (north of the latitude of Cairns, 17° S.) has been as yet untested. According to Stanley (1928*b*), it is only in this warmer northern half that we would expect to encounter Pleistocene reefs. He believed that most of the reefs that we see today have grown up in postglacial times. Thus a hole put down some distance north of Cairns should disclose a very different section from those encountered in the two southern ones. This finds support in Haddon's discovery of blocks of dolomite in the volcanics of the Murray Islands (1894), as Hedley and Taylor remarked (1908): "It seems a fair deduction to conclude that at this spot sunken reefs exist below the 100 fathom level."

The evidence of the Great Barrier Reef borings offers no support to one of the original tenets of Daly's glacial-control theory (1910) which required the reefs to grow up from a shelf that had been planed down by Pleistocene marine erosion. Daly specifically mentioned the Australian continental shelf (1910, p. 305). However, as pointed out by Ladd and Hoffmeister (1936), this control was later recognized by Daly to be of minor importance and possibly not effective at all. The foundation of the reefs in question is a thick and unconsolidated layer of youthful terrigenous sediments. These must have been laid down on the continental shelf by normal depositional processes and distributed evenly over a wide area by currents. The period of sedimentation must have been long and the rate of sedimentation slow, as shown by the glauconites and the absence of fresh

feldspars. A fairly shallow, current-swept shelf is shown by the fossils and the absence of clays. The noncalcareous nature of this material may be taken, perhaps, as evidence of cold-water (Pleistocene) deposition.

The evidence of the boreholes does not positively confirm or deny, for the Queensland reefs at least, the various antecedent abrasional-platform hypotheses of Wharton, Andrews, Vaughan, Hoffmeister and Ladd, and others. On geomorphologic and tectonic grounds, we are persuaded of the general correctness of this antecedent-platform concept, but eustatic influences (Daly) and a degree of subsidence (Darwin) are also clearly evident.

CONCLUSIONS

1. Reefs of the shelf have been divided into four principal types based on form: fringing, barrier, atoll, and platform. These terms have no genetic implication.

2. Five main types of coral island are distinguished: unvegetated sand cay; stable, vegetated sand cay; shingle cay; sand cay with shingle ridges or "ramparts" on the reef platform; coral islands with partly eroded core of older emergent coral limestone (see also conclusions 5 and 6).

3. The shape of reefs is controlled principally by winds, waves, and currents. Several stages of reef evolution are recognized. Stage 1 is the initial reef patch; stage 2, crescentic or horseshoe patch. Following these stages the larger and smaller types develop differently. Stage 3*A* (larger) is the semiatoll; stage 3*B* (smaller) is a horsehoof-shaped reef. Stage 4*A* of the larger reef is a complete atoll, which is followed by stage 5*A*, an infilled platform reef. Stage 4*B*, or end-stage of the smaller reef, is a small, oval, reef platform.

The reef shape is influenced also by old geomorphologic and structural features. A fringing reef, for example, may be initiated along a shore line when the sea level is lower than today. Reefs may follow a submerged fault scarp or ranges of drowned hills. The Outer Barrier Reef (in the northern sector) exhibits both structural and geomorphological controls, but the curving wings at the extremities of each ribbon-like section of the reef are due to wind and current control.

4. Coral islands on the Australian shelves may be explained in either of two ways: (*a*) By accumulation of sand or coral shingle under wave action, to form cays. The lighter sand forms a bank at the leeward extremity of the reef-flat and eventually a sand cay, whereas coral shingle is thrown up only a short distance by waves on the windward side to form shingle beach ridges, ramparts, or, in a group, a shingle cay. (*b*) By a eustatic drop in sea level, exposing older coral reefs. Since the mid-Recent high sea level, about 10 feet above the present level, some four thousand years ago, there have been three successive drops in sea level, the first to 5 feet, the second to 2 feet, and the third to the present stand. Coral reefs formed during each of these stable periods now form the nucleus of many coral islands. Still older coral limestones (Pleistocene) are found in the cores of the Abrolhos Islands.

5. Coral islands due to accumulation are generally rounded or oval in shape, "streamlined," even though they periodically undergo erosion. Those produced by a eustatic drop in sea level possess a scalloped, irregular outline and generally undergo rapid erosion except where they are protected by new shingle ridges, etc.

6. "Negroheads," isolated coral boulders of large dimensions on reef margins,

have been thrown up in some places from the reef margin by storms, and in others they are due to erosion *in situ* of emerged coral limestone terraces and islets.

7. The distribution of the coral reefs around the Australian continental shelves cannot be explained by ecologic factors alone. Except in shallow, favored areas, coral reefs are not growing up from the floor of the shelf today. The platform, atoll, and barrier types of reefs, therefore, require some change of sea level. The changes may have been the eustatic swing that occurred during the Pleistocene and Recent epochs; but those reefs rising from depths exceeding 300 feet must be related to tectonic movements of the shelf itself. An adequate account of reef distribution and evolution requires consideration of ecologic factors, eustatic factors, and tectonic factors—altogether and *not* singly.

8. Fringing reefs can be explained in ecologic terms. They are common on northern shores but are rare elsewhere; on the southern part of the Queensland Shelf this is probably due to the excessive rainfall and sediment, around the Gulf of Carpentaria because of rivers and sediment, along the Eighty Mile Beach and many sections of the western coast because of sandy bottom, lee shore, and poor nutrients. Most fringing reefs show an inner, raised platform, a mid-Recent eustatic terrace, commonly covered with sediment and in part with mangroves.

9. The inner reefs (small patch reefs, island reefs, platform reefs, and the few shelf atolls of the Great Barrier Reef lagoon) rise from a shelf terrace submerged 15 or 25 fathoms and therefore possess a linear arrangement. Broad open channels are found in the Barrier Reef lagoon where there are no inner reefs. The latter probably evolved from former discontinuous fringing reefs during the postglacial eustatic rises of sea level, because the water is generally too deep in the central part of the shelf for reefs to have grown up from the floor under present ecologic conditions. It is unlikely that the Great Barrier Reef lagoon will be blocked by coral growth unless sedimentation considerably reduces its depth.

10. The outer-barrier reefs of Queensland (northern sector) rise from depths of over 100 fathoms. Because this exceeds the sum of the present ecologic limits of reef growth (here about 20 fathoms) and the 50 fathoms allowed for maximum Pleistocene eustatic lowering, tectonic subsidence must have occurred. The central sector (Cairns-Townsville) shows a drowned outer barrier which suggests very rapid subsidence locally because the postglacial rise elsewhere failed to drown long sections of reef. The southern sector rises inside the 100-fathom line in all places. No coral growth may have occurred here during much of the Pleistocene because of the adverse climate. Only in the last warm stages of the Pleistocene was coral growth initiated.

11. The great atoll reefs of the Coral Sea Platform, which adjoin the outer barrier, show evidence of tectonic subsidence to as much as 6,000 feet.

12. On the Sahul and North-West shelves, several large atolls occur near the shelf margins in 200-300 fathoms of water. This depth greatly exceeds the usual estimates for eustatic rise of sea level. This fact plus the atoll forms, regional low level of the shelf edge, and various structural evidence suggest tectonic subsidence.

13. The complex atoll groups of the Houtman Abrolhos Islands on the west-coast shelf rise only from a 25-fathom shelf and require no hypothesis of subsidence. They disclose a core of old coral

reef and lagoon limestones, which was heavily dissected by karst erosion during the late Pleistocene and veneered by reefs and coral shingle of the various mid-Recent stillstands and of the present day.

14. In the Great Barrier Reef lagoon sedimentation is extremely rapid. The deposits are mainly terrigenous, although big lenses of coral and coral debris are intercalated in them. In the borings at Michaelmas Cay and Heron Island, "Recent" reef corals (with other "Recent" fossils) go down over 400 and 500 feet, respectively. Because the present shelf floor is only 90–150 feet below the surface, an accumulation of 300–400 feet of sediment during the last hundred thousand years seems probable. The absence of submarine valleys across the shelf is another indication of rapid filling by postglacial sediments. The rapid accumulation is attributed to the retaining effects of the outer-reef barrier.

15. Slow or even suspended terrigenous sedimentation characterizes the open-shelf areas (Arafura, Sahul, North-West, and West Coast shelves). Glauconites are common; coral materials are rare. The small outcrop of the granite basement on the Aroe Islands, near the outer edge of the Arafura shelf, suggests only a thin covering of Tertiary and Quaternary sediments.

16. Even in the regions of most vigorous reef growth, the proportion of reef sediments to terrigenous material is small. In the reefs proper, the proportion of true coral grown *in situ* to coraligenous sedimentary debris and other associated organogenic material is small. The so-called "coral" sands generally contain less than one-third coral debris.

17. The calcium carbonate muds in the borings of both lagoons are in part fine clastics and in part true precipitates.

18. The massive coral growths are a rough framework in and around which loose and uncompacted clastic calcareous sediments accumulate. In fringing reefs this material accumulates on the seaward side, in barrier reefs mainly on the inner side; and in atolls it fills the interior and accumulates in a cone to leeward of the reef. Most platform reefs probably originated in the same way as barriers and atolls did during progressive submergence. The circular or ovoid ones exhibit what appears to be a filled-in lagoonal core.

19. Sedimentation, modern or Pleistocene, influences the distribution and growth of reefs in only a secondary way. Reef growth is greatest in the Great Barrier Reef lagoon, where sedimentation is greatest, whereas in areas of minimum sedimentation, like the Sahul Shelf, coral reefs are comparatively rare.

20. Great Barrier Reef borings show not less than 200 feet of loose glauconitic quartz sands (Pleistocene) beneath the 400–500 feet of "Recent" coral material.

21. Calcareous beach rock, an intertidal deposit, occurs in the deepest boring, some 700 feet below sea level. Foraminifera and Mollusca in these sands are all of living types and cannot be older than Pleistocene.

22. Tectonic subsidence in the Great Barrier Reef area is probable. If the beach rock at 700 feet (Heron Island boring) was formed during a Pleistocene low sea level (minimum generally taken as 300 feet), then 400 feet of subsidence has taken place since. If the 400–450-foot level in the Michaelmas Cay boring corresponds to the 450–500-foot level in the Heron Island hole, then subsidence has been 50–100 feet greater in the latter.

23. The borings of Florida and other continental shelves also show terrigenous (mainly quartz) sands, likewise overlain by several hundred feet of reef rock, loose

coralline debris, and foraminiferal sands. The borings on Funafuti, Bikini, and other oceanic atolls show less similarity. Little or no terrigenous material and thicker Quaternary coral deposits suggest more rapid subsidence.

24. No dolomite is found in any Australian reef borings or emerged coral reefs. Submarine dolomite formation is now thought to take place in closed lagoons under reducing conditions and increased CO_2 pressure, an environment almost totally absent from the Australian region. The only dolomite in the Quaternary rocks of Australia consists of blocks enclosed in volcanics on Murray Island (Haddon, 1894). Perhaps the northern section of the Great Barrier Reef may rest on a basement older than, or different from, that of the southern sector.

25. Magnesium carbonate enrichment, organic and inorganic, takes place in the intertidal belt. In borings on the Barrier Reef and at Bikini, Funafuti, etc., a sharp rise in $MgCO_3$ from a normal of about 1–5 per cent to 5–16 per cent is marked by beach-rock layers, indicating stillstands, either tectonic or eustatic. Low $MgCO_3$ indicates rapid change of sea level, whereas higher figures suggest relative stability.

26. Aragonite is found in the Great Barrier Reef borings, as in Funafuti, in the upper 50–100 feet, below which it is almost completely leached out and converted to calcite. This leaching is attributed to subaerial exposure during the Pleistocene lowering of sea level. At 400 feet in Michaelmas Cay and 500 feet in the Heron Island boring, the aragonite reappears, suggesting that the sections below these levels were not exposed to late Pleistocene subaerial erosion. The failure of aragonite to reappear in the Funafuti boring implies that, in this rapidly subsiding atoll, the whole section was exposed progressively to repeated exposures during the Pleistocene lows.

27. Neither Great Barrier Reef boring reached hard basement. An antecedent platform is probable. Its character is revealed on various continental islands, where granite and metamorphic rocks of Paleozoic or greater age are exposed. The basement is generally deeply buried by sediment, not less than 700 feet at Heron Island, and is not well known.

28. Existence of a submerged plane of erosion is demonstrated by cliffed and steep coasts, drowned drainage systems, etc. Borings in the alluvial coastal plain of the mainland show the hard floor over 100 feet below sea level in Queensland and as much as 250 feet below in New South Wales. The cutting of this erosion plane in hard rocks required a considerable period of time and could not have been done during the short periods of low sea level at the Pleistocene glacial epochs. It is correlated with the great Tertiary peneplanation of eastern Australia, which was terminated by epeirogenic warping and fracturing in late Tertiary times. During this "Kosciusko" phase (of Andrews), the eastern highlands were elevated *en bloc*, and the adjacent areas to the east subsided along major normal faults.

29. The outer margin of the continental shelf is apparently down-faulted along the entire eastern seaboard of Australia. This is demonstrated by the abruptness of the slope, by the almost rectilinear pattern, and by such structural and paleogeographic evidence as the truncation of former geosynclines, fresh-water basins, and the total absence of marine Tertiary sediment.

30. The subsided land mass ("Tasmantis" of Sussmilch and David) is partly represented by the floor of the Tasman

Sea (in the south), but in the north, opposite the Queensland Shelf, it is marked by a northward-tilted block, the Coral Sea Platform, separated from the shelf by a deep graben, the Queensland Trench, while on the north it is cut off from the New Guinea–New Hebrides orogenic belt by a 2,000-fathom-deep extension of the Papuan Geosyncline, the Papuan Trough.

31. Because the continental shelf of Queensland is structurally identical with that of New South Wales, where there are no corals, the Queensland reefs cannot be thicker than the depth below sea level of the shelf basement—not that of the superficial sediments, which the borings have shown may exceed 600 feet in thickness. Borings have not shown reefs here to exceed 500 feet in thickness.

32. The glauconitic quartz sands of the lower 200 feet of the Great Barrier Reef borings are probably Pleistocene, formed in a period of coral depopulation. This conclusion is supported by living species throughout the fossil record in the borings.

33. The thickness and character of the Pleistocene coral reefs, if any, in the more tropical belts of Australia are unknown.

34. The hard basement of the northern shelf areas is thought to be fairly shallow in places (e.g., small outcrop of granite in Aroe Islands) but is probably deeper off the basins of the mainland. Subsidence seems to be in progress opposite the mobile East Indian arcs, resulting in atolls and a drowned barrier reef which may exceed 1,500 feet in thickness.

35. The hard basement of the western shelf also appears to be shallow. The Abrolhos Islands apparently stand on the margin of an uptilted fault block of pre-Cambrian rocks.

36. From the stratigraphic viewpoint, it is seen that coral reefs may grow even in so-called "stable" shelf environments. Shallow, thin-bedded reefs develop laterally along immense stretches of coast line under conditions of tectonic and eustatic stability, whereas poorly bedded and massive reefs, of limited lateral extent but of considerable thickness, develop under unstable tectonic and eustatic conditions.

37. The Australian shelf reefs require the utilization, at least of some parts, of (*a*) the subsidence theory promulgated by Darwin, Dana, and Davis, (*b*) the antecedent-platform theory as set forth by Wharton, Agassiz, Andrews, Vaughan, Hoffmeister and Ladd, and (*c*) the glacial-control theory of Penck, Daly, and others.

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REFERENCES CITED

ADMIRALTY CHARTS, Australia Pilot, vols. 3–5, Hydrographic Department, London, H.M. Stationery Office.

AGASSIZ, A. (1894) A reconnaissance of the Bahamas and of the elevated reefs of Cuba: Harvard Mus. Comp. Zoology Bull. 26, pp. 1–203.

——— (1895) A visit to the Bermudas in March, 1894: *ibid.*, pp. 205–281.

——— (1898) A visit to the Great Barrier Reef of Australia in the steamer "Croyden" during April and May, 1896: *ibid.*, vol. 28, pp. 93–148.

——— (1903) On the formation of barrier reefs and of the different types of atolls: Royal Soc. London Proc., vol. 71, pp. 412–414.

AHLMAN, H. W. (1949) The present climatic fluctuations: Geog. Jour., vol. 112, pp. 165–195.

ANDREWS, E. C. (1902) A preliminary note on the geology of the Queensland coast, with reference to the geography of the Queensland and New South Wales plateaux: Linnean Soc. New South Wales Proc., vol. 27, pp. 146–185.

——— (1910) Geographical unity of eastern Australia in late and post-Tertiary time: Royal Soc. New South Wales Jour. and Proc., vol. 44, pp. 420–480.

——— (1922) A contribution to the hypothesis of coral reef formations (presidential address): *ibid.*, vol. 56, pp. 10–38.

AUROUSSEAU, M., and BUDGE, E. A. (1921) The terraces of the Swan and Helena rivers and their bearing on Recent displacement of the strandline: Royal Soc. Western Australia Jour. and Proc., vol. 7, pp. 24–43.

BAKER, J. R. (1925) A coral reef in the New Hebrides: Zool. Soc. London Proc., pp. 1007–1019.

BASEDOW, H. (1916) Physical geography and geology of the Western Rivers' District: Royal Geog. Soc. Australia Proc. (S. A. Branch), vol. 16 (1914/15), pp. 143–217.

——— (1918) Narrative of an expedition of exploration in north-western Australia: *ibid.*, vol. 18 (1916/17), pp. 105–295. (See especially "A synopsis of the geology . . . ," pp. 240–249.)

BASSET-SMITH, P. W. (1899) On the formation of the coral reefs on the N.W. coast of Australia: Zool. Soc. London Proc., pp. 157–159.

BEASLEY, A. W. (1947) The place of black sand seams in the physiographic history of the South Coast region, Queensland: Australian Jour. Sci., vol. 9, pp. 208–210.

——— (1948) Heavy mineral beach sands of southern Queensland: Royal Soc. Queensland Proc., vol. 59, pp. 109–140.

BLACK, M. A. (1933) The precipitation of calcium carbonate on the Great Bahama Bank: Geol. Mag., vol. 70, pp. 455–466.

BOUCHART, J. (1938) La marge continentale. Essai sue les regressions et transgressions marines: Soc. géol. France Bull. 8, 5th ser., pp. 393–474.

BRANNER, J. C. (1905) Stone reefs on the north-east coast of Brazil: Geol. Soc. America Bull. 16, pp. 1–13.

BROWN, H. Y. L. (1906) Explorations made by the Government Geologist, 1905: South Australia Parliamentary Paper 55.

——— (1908) Geological reconnaissance from Van Dieman Gulf to the McArthur River, etc.: *ibid.*, no. 25.

BROWNE, W. R. (1945) An attempted post-Tertiary chronology for Australia: Linnean Soc. New South Wales Proc., vol. 70, pp. v–xxiv.

BRYAN, W. H. (1928) The Queensland continental shelf: Great Barrier Reef Comm. Repts., vol. 2, pp. 58–69.

——— (1936) The supposed deepening of the sea floor off Breaksea Spit: *ibid.*, vol. 4, pt. 2, pp. 45–50.

——— (1944) The relationship of the Australian continent to the Pacific Ocean—now and in the past: Royal Soc. New South Wales Jour. and Proc., vol. 78, pp. 42–62.

CADELL, H. M. (1899) Some geological features of the coast of Western Australia: Edinburgh Geol. Soc. Trans., vol. 7 (1895/6), pp. 174–182.

CHAPMAN, F. (1931) A report on samples obtained by boring into Michaelmas Reef, about 22 miles N.E. of Cairns, Queensland: Great Barrier Reef Comm. Repts., vol. 3, pp. 32–42.

——— (1944) The Foraminifera of the Funafuti boring: Annals and Mag. Nat. History, 11th ser., vol. 11, no. 74, IX, pp. 98–110.

CLAPP, F. G. (1926) Oil prospects of the Desert Basin of Western Australia: Am. Assoc. Petroleum Geologists Bull. 10, pp. 1118–1135.

CLARKE, F. W., and WHEELER, W. C. (1922) The inorganic constituents of marine invertebrates: U.S. Geol. Survey Prof. Paper 124, 2d ed.

CONDIT, D. D. (1935) Oil possibilities in Northwest District, Western Australia: Econ. Geology, vol. 30, pp. 860–878.

COOKE, C. W. (1945) Geology of Florida: Florida Geol. Survey Bull. 29, 339 pp.

COTTON, C. A. (1949) A review of the tectonic relief of Australia: Jour. Geology, vol. 57, pp. 280–296.

COTTON, L. A. (1947) The pulse of the Pacific: Royal Soc. New South Wales Jour. and Proc., vol. 80, pp. 41–76.

CROSSLAND, C. (1902) The coral reefs of Zanzibar: Cambridge Philos. Soc. Proc., vol. 11, pp. 493–503.

——— (1907) Note on the formation of the shore-cliff near Alexandria: Linnean Soc. London Jour., vol. 31, pp. 10–13.

——— (1913) Desert and water gardens of the Red Sea, Cambridge, University Press, 158 pp.

——— (1928*a*) Coral reefs of Tahiti, Moorea, and Rarotonga: Linnean Soc. London Jour., vol. 36, pp. 577–620.

——— (1928*b*) Notes on the ecology of the reef-builders of Tahiti: Zool. Soc. London Proc., pp. 717–735.

——— (1931) The coral reefs of Tahiti compared with the Great Barrier reefs: Geog. Jour., vol. 77, pp. 395–396.

CULLIS, C. G. (1904) The mineralogical changes observed in the cores of the Funafuti borings; in:

The atoll of Funafuti, Royal Soc., Coral Reef Comm. Rept., pp. 392–420.

CUSHMAN, J. A. (1942) A report on samples obtained by the boring at Heron Island, Great Barrier Reef, Australia: Great Barrier Reef Comm. Repts., vol. 5, pp. 112–119.

DAKIN, W. J. (1919) The Percy Sladen Trust Expeditions to the Abrolhos Islands (Indian Ocean): Linnean Soc. London Jour. (Zoology), vol. 34, pp. 127–180.

DALY, R. A. (1910) Pleistocene glaciation and the coral reef problem: Am. Jour. Sci., 4th ser., vol. 30, pp. 297–308.

——— (1915) The glacial-control theory of coral reefs: Am. Acad. Arts Sci. Proc., vol. 51, pp. 155–251.

——— (1920) A recent world-wide sinking of ocean-level: Geol. Mag., vol. 57, pp. 246–261.

——— (1925) Pleistocene changes of level: Am. Jour. Sci., 5th ser., vol. 10, pp. 281–313.

——— (1929) Swinging sea-level of the ice age: Geol. Soc. America Bull. 40, pp. 721–734.

——— (1934) The changing world of the ice age, New Haven, Yale University Press, 271 pp.

——— (1948) Coral reefs—a review: Am. Jour. Sci., vol. 246, pp. 193–207.

DANA, J. D. (1849) Geology, U.S. Exploring Expedition 1838/42, Philadelphia, vol. 10.

——— (1873) Corals and coral islands, New York, Dodd, Mead & Co., 440 pp.

DANEŠ, J. V. (1911) On the physiography of north-eastern Australia: Soc. royale sci. Bohème Proc., Prague, 2d ser., 18 pp.

——— (1912) La région des rivières Barron et Russel (Queensland): Annales de géographie, vol. 21, pp. 344–363.

DARWIN, C. (1842) The structure and distribution of coral reefs, London, 214 pp.

DAVID, T. W. E. (1911) Notes on some of the chief tectonic lines of Australia: Royal Soc. New South Wales Jour. and Proc., vol. 45, pp. 4–60.

——— and HALLIGAN, G. H. (1908) Evidence of Recent submergence of coast at Narrabeen: Royal Soc. New South Wales Jour. and Proc., vol. 42, pp. 229–237.

———, ———, and FINCK, A. E. (1904) Report on dredging at Funafuti; in: The atoll of Funafuti, Royal Soc., Coral Reef Comm. Rept., pp. 151–159.

——— and SWEET, G. (1904) The geology of Funafuti; in: The atoll of Funafuti, Royal Soc., Coral Reef Comm. Rept., pp. 61–124.

DAVIS, W. M. (1917) The Great Barrier Reef of Australia: Am. Jour. Sci., 4th ser., vol. 44, pp. 339–350.

——— (1928) The coral reef problem: Am. Geog. Soc. Special Pub. 9.

——— (1934) Gardiner on coral reefs and atolls—a discussion: Jour. Geology, vol. 42, pp. 200–217.

EDGELL, J. A. (1928*a*) Some remarks on coral formations: Great Barrier Reef Comm. Repts., vol. 2, pp. 52–56.

——— (1928*b*) Changes at Mast Head Island: *ibid.*, p. 57.

FAIRBRIDGE, R. W. (1946) Coarse sediments on the edge of the continental shelf: Am. Jour. Sci., vol. 245, pp. 146–153.

——— (1947) A contemporary eustatic rise in sea-level? Geog. Jour., vol. 109, p. 157.

——— (1948*a*) Notes on the geomorphology of the Pelsart Group of the Houtman's Abrolhos Islands: Royal Soc. Western Australia Jour., vol. 33 (for 1946/1947), pp. 1–43.

——— (1948*b*) Discoveries in the Timor Sea, North-West Australia: Royal Australian Hist. Soc. Jour. and Proc., vol. 34, pp. 193–218.

——— (1948*c*) The juvenility of the Indian Ocean: Scope (Univ. Western Australia Jour. Sci.), vol. 1, no. 3, pp. 29–35.

——— (1950*a*) Multiple stands of the sea in post-glacial times: Seventh Pacific Sci. Cong. Proc. (New Zealand, 1949). (In press.)

——— (1950*b*) Marine erosion: *ibid.* (In press.)

——— (1950*c*) The geology and geomorphology of Point Peron, Western Australia: Royal Soc. Western Australia Jour., vol. 34, no. 3, pp. 35–72.

——— (1950*d*) The Sahul Shelf, northern Australia: its structure and geological relationships, *ibid.*, vol. 36. (In press.)

——— (1950*e*) Landslide patterns on oceanic volcanoes and atolls: Geog. Jour., vol. 115, pp. 84–88.

——— and TEICHERT, C. (1947) The rampart system at Low Isles, 1928–45: Great Barrier Reef Comm. Repts., vol. 6, pp. 1–6.

——— — ——— (1948) The Low Isles of the Great Barrier Reef: a new analysis: Geog. Jour., vol. 111, pp. 67–88.

FRANCIS-BŒUF, C. (1938) La problème du plateau continental et des vallées sous-marines: Rev. géographie phys. et géologie dynamique, vol. 11, p. 254.

FRYER, J. C. F. (1910) The south-west Indian Ocean: Geog. Jour., vol. 36, pp. 251–271.

——— (1911) The structure and formation of Aldabra and neighbouring islands: Linnean Soc. London Trans., 2d ser., vol. 14, pp. 397–442.

GARDINER, J. S. (1898) The coral reefs of Funafuti, Rotuma, and Fiji, together with some notes on the structure and formation of coral reefs in general: Cambridge Philos. Soc. Proc., vol. 9, pp. 417–503.

——— (1903–6) The fauna and geography of Maldive and Laccadive archipelagos (the Maldive and Laccadive groups, with notes on other coral formations in the Indian Ocean), 2 vols., Cambridge, Cambridge University Press.

——— (1906) The Indian Ocean. Being results largely based on the work of the Percy Sladen Expedition in H.M.S. Sealark, Com. B. T.

Somerville, 1905: Geog. Jour., vol. 28, pp. 313–332; 454–471.

——— (1915) Submarine slopes: *ibid.*, vol. 45, pp. 202–219.

——— (1930) Studies in coral reefs: Harvard Coll. Mus. Comp. Zoology Bull. 71, pp. 1–16.

——— (1931*a*) Photosynthesis and solution in formation of coral reefs: Linnean Soc. London Proc., 1930/31, pp. 65–71.

——— (1931*b*) Coral reefs and atolls, London, Macmillan & Co., Ltd.

——— (1936) The reefs of the western Indian Ocean. I. Chagos Archipelago. II. The Mascarene region: Linnean Soc. London Trans. (Zoology), vol. 19, pp. 393–436.

——— and COOPER, C. F. (1907) Description of the Expedition: Linnean Soc. London Trans. (Zoology), 2d ser., vol. 12, pp. 1–56; 111–175.

GREGORY, J. W. (1912) The structural and petrographical classification of coast-types: Scientia, vol. 2, pp. 36–63.

——— (1930) The geological history of the Pacific Ocean (anniversary address): Geol. Soc. London Quart. Jour., vol. 86, pp. lxxii–cxxvi.

GUPPY, H. B. (1884) Suggestions as to the mode of formation of barrier reefs in Bougainville straits, Solomon group: Linnean Soc. New South Wales Proc., vol. 9, pp. 949–959.

——— (1887) The Solomon Islands, London, S. Sonnenschein, Lowrey & Co.

——— (1888) A criticism of the theory of subsidence as affecting coral reefs: Scottish Geog. Mag., vol. 4, pp. 121–137.

——— (1890) The origin of coral reefs: Victoria Inst. Jour. (Philos. Soc. Great Britain), vol. 23, pp. 51–68.

GUTENBERG, B. (1941) Changes in the sea-level, post-glacial uplift, and mobility of the earth's interior: Geol. Soc. America Bull. 52, pp. 721–772.

HADDON, A. C. (1935) General ethnography, Cambridge Anthropological Expedition to Torres Straits Repts., vol. 1 (Geography and Geology, pp. 19–36).

———, SOLLAS, W. J., and COLE, G. A. J. (1894) On the geology of the Torres Straits: Royal Irish Acad. Trans., vol. 30, pp. 419–476.

HANZAWA, S. (1940) Micropaleontological studies of drill cores from a deep well in Kita-Daito-Jima: H. Yabe Jubilee Pub., vol. 2, pp. 755–802.

HARDMAN, E. T. (1884) Report on the geology of the Kimberley district, Western Australia, Perth, Government Printer.

HARTMEYER, R. (1907) Die Fauna Südwest-Australiens (Reisebericht, 2), vol. 1, Jena, pp. 59–108.

HATCH, F. H.; RASTALL, R. H.; and BLACK, M. (1938) The petrology of the sedimentary rocks, 3d ed., London, Allen & Unwin, 383 pp.

HEDLEY, C. (1906) The Mollusca of Mast Head Reef, Capricorn Group, Queensland: Linnean Soc. New South Wales Proc., vol. 31, pp. 453–479.

——— (1910) The submarine slope of New South Wales: *ibid.*, vol. 35, pp. 9–22.

——— (1925*a*) The natural destruction of a coral reef: Great Barrier Reef Comm. Repts., vol. 1, pp. 35–40.

——— (1925*b*) A raised beach in the North Barnard Islands: *ibid.*, pp. 61–62.

——— (1925*c*) The surface temperature of Moreton Bay: *ibid.*, pp. 149–150.

——— (1925*d*) Report of the Scientific Director for 1924: *ibid.*, pp. 157–160.

——— (1926) Recent studies on the Great Barrier Reef of Queensland: Am. Jour. Sci., 5th ser., vol. 11, pp. 187–193.

——— and TAYLOR, T. G. (1908) Coral reefs of the Great Barrier, Queensland: Australian Assoc. Adv. Sci. Rept., vol. 11, pp. 397–413.

HINDE, G. J. (1904) Report on the materials from the borings at the Funafuti Atoll; in: The atoll of Funafuti, Royal Soc., Coral Reef Comm. Rept., pp. 186–361.

HOFFMEISTER, J. E. (1932) Geology of Eua, Tonga: Bernice P. Bishop Mus. Bull. 96.

——— and LADD, H. S. (1944) The antecedent-platform theory: Jour. Geology, vol. 52, pp. 388–402.

HOVEY, E. O. (1896) Notes on the artesian well sunk at Key West, Fla., in 1895: Harvard Mus. Comp. Zoology Bull. 38, pp. 65–91.

IREDALE, T. (1942) Report on molluscan content of Heron Island reef boring samples: Great Barrier Reef Comm. Repts., vol. 5, pp. 120–122.

JACK, R. L., and ETHERIDGE, R. (1892) The geology and palaeontology of Queensland and New Guinea, Brisbane, Government Printer, 768 pp.

JACKSON, C. F. V. (1902) Report on a visit to the west coast of the Cape York Peninsula and some islands of the Gulf of Carpentaria: Queensland Geol. Survey Rept. 180, 27 pp.

JARDINE, F. (1925) The development and significance of benches in the littoral of eastern Australia: Great Barrier Reef Comm. Repts., vol. 1, pp. 111–130.

——— (1928*a*) The topography of the Townsville littoral: *ibid.*, vol. 2, pp. 70–87.

——— (1928*b*) The Broadsound drainage in relation to the Fitzroy River: *ibid.*, pp. 88–92.

——— (1928*c*) Bramble Cay, Torres Strait—geological notes: *ibid.*, pp. 93–100.

——— (1928*d*) Darnley Island—geological and topographical notes: *ibid.*, pp. 101–109.

JENSEN, H. I. (1911) The building of eastern Australia: Royal Soc. Queensland Proc., vol. 23, pp. 149–198.

——— (1914) Geological report on the Darwin mining districts, McArthur River district; and the Barkly tableland: Northern Territory Dept. External Affairs Bull. 10, 38 pp.

JESSEN, O. (1943) Die Randschwellen der Kontinente: Petermanns Mitt., Ergänzungsheft 241.

JOUBIN, L. (1912) Bancs et récifs de coraux (Madrépores)—texts et carte: Annales de Inst. Océanographie, vol. 4, no. 2.

JUDD, J. W. (1904) The chemical examination of the materials from Funafuti; in: The atoll of Funafuti, Royal Soc., Coral Reef Comm. Rept., pp. 362–389.

JUKES, J. B. (1847) Narrative of the surveying voyage of H.M.S. "Fly," London, 2 vols.

JUTSON, J. T. (1934) The physiography (geomorphology) of Western Australia: Western Australia Geol. Survey Bull. 95, 2d rev. ed.

KREMPF, A. (1927) La forme des récifs coralliens et le régime des vents alternants: Indochine Travaux Service Océanographie, Mem. 2, 33 pp.

——— (1930) About the shape of the coral reefs and the system of the alternating winds: Fourth Pacific Sci. Cong. Proc., vol. 2A, pp. 477–480.

KUENEN, PH. H. (1933) Geology of coral reefs; in: The Snellius Expedition, vol. 5, Geological results, pt. 2, 125 pp.

LADD, H. S., and HOFFMEISTER, J. E. (1927) Recent negative shift in the strand line in Fiji and Tonga: Jour. Geology, vol. 35, pp. 542–556.

——— — ——— (1936) A criticism of the glacial-control theory: *ibid.*, vol. 44, pp. 74–92.

———, TRACEY, J. I., and LILL, G. G. (1948) Drilling on Bikini Atoll, Marshall Islands: Science, vol. 107, pp. 51–55.

LE CONTE, J. (1856) On the agency of the Gulf Stream in the formation of the peninsula and keys of Florida: Am. Jour. Sci., 2d ser., vol. 23, pp. 46–60.

LENDENFELD, R. VON (1902) Das grosse australische Wallriffe: Zeitschr. Geographie, vol. 8, pp. 369–379.

LENOX-CONYNGHAM, G., and POTTS, F. A. (1925) The Great Barrier Reef: Geog. Jour., vol. 65, pp. 314–334.

LISTER, J. J. (1891) Notes on the geology of the Tonga Islands: Geol. Soc. London Quart. Jour., vol. 47, pp. 590–617.

MACFADYEN, W. A. (1930) The undercutting of coral reef limestone on the coasts of some islands in the Red Sea: Geog. Jour., vol. 75, pp. 27–34.

MACGILLIVRAY, T. (1852) Narrative of a voyage of H. M. S. *Rattlesnake*, 1846–50, London, 2 vols.

MANTON, S. M. (1935) Ecological surveys of coral reefs: Great Barrier Reef Expedition, British Mus. Nat. History, vol. 3, pp. 273–312.

MARSHALL, M. A.; RICHARDS, H. C.; and WALKOM, A. B. (1925) Recent emergence at Holbourne Island, Great Barrier Reef: Great Barrier Reef Comm. Repts., vol. 1, no. 2, pp. 29–34.

MARSHALL, P. (1910) Ocean contours and earth movements in the southwest Pacific: Australian Assoc. Adv. Sci. Rept., vol. 10, pp. 432–450.

——— (1930) Geology of Rarotonga and Atiu: Bernice P. Bishop Mus. Bull. 72.

——— (1931) Coral reefs—rough-water and calm-water types: Great Barrier Reef Comm. Repts., vol. 3, pp. 64–72.

——— (1933) Stability of lands in the South-west Pacific: Australian and New Zealand Assoc. Adv. Sci. Rept., vol. 21, pp. 399–411.

MARSHALL, S. M., and ORR, A. P. (1931) Sedimentation on Low Isles reef and its relation to coral growth: Great Barrier Reef Expedition, 1928–29, vol. 1, no. 5, pp. 93–133.

MATSON, G. C., and SANFORD, S. (1913) Geology and ground waters of Florida: U.S. Geol. Survey Water Supply Paper 319.

MAYER, A. G. (1915) Ecology of the Murray Island coral-reef: Nat. Acad. Sci. Proc., vol. 1, pp. 211–214.

——— (1918) Ecology of the Murray Island coral reef: Carnegie Inst. Washington Pub. 213. Dept. Marine Biology Papers, vol. 9, pp. 1–48.

MAYOR, A. G. (1924) Growth-rate of Samoan corals: Carnegie Inst. Washington Pub. 340. Dept. Marine Biology Papers, vol. 19, pp. 51–72.

MOLENGRAAFF, G. A. F. (1917) The coral reef problem and isostasy: K. Akad. Wetensch. Amsterdam Proc., vol. 19, pp. 610–627.

——— (1930) The coral reefs of the East Indian Archipelago, their distribution and mode of development: Fourth Pacific Sci. Cong. Proc., vol. 2, A and B, pp. 55–99, 989–1021.

MOORHOUSE, F. W. (1936) The cyclone of 1934 and its effects on Low Isles: Great Barrier Reef Comm. Repts., vol. 4, pp. 37–44.

MURRAY, J. (1880) On the structure and origin of coral reefs and islands: Royal Soc. Edinburgh Proc., vol. 10, pp. 505–518.

——— and IRVINE, R. (1891) On coral reefs and other carbonate of lime formations in modern seas: Royal Soc. Edinburgh Proc. (A), vol. 17, pp. 79–109.

——— and RENARD, A. F. (1891) Deep-sea deposits, Rept. on the . . . voyage of H.M.S. "Challenger" 1873–76, 525 pp.

NELSON, R. J. (1837) On the geology of the Bermudas: Geol. Soc. London Trans., 2d ser., vol. 5, pp. 103–123.

——— (1853) On the geology of the Bahamas, and on coral formations generally: Geol. Soc. London Quart. Jour., vol. 9, pp. 200–215.

NUGENT, L. E. (1946) Coral reefs in the Gilbert, Marshall and Caroline islands: Geol. Soc. America Bull. 57, pp. 735–780.

——— (1948) Elevated phosphate islands in Micronesia: *ibid.*, vol. 59, pp. 977–994.

ORR, A. P. (1933) Physical and chemical conditions in the sea in the neighbourhood of the Great Barrier Reef: Great Barrier Reef Expedition, 1928–29, British Mus. Nat. History Sci. Repts., vol. 2, no. 3, pp. 37–86.

——— and MOORHOUSE, F. W. (1933a) Variations in some physical and chemical conditions on and near Low Isles reef: Great Barrier Reef Expedi-

tion, 1928–29, British Mus. Nat. History Sci. Repts., vol. 2, no. 4(a), pp. 87–98.

——— —— ——— (1933*b*) Physical and chemical conditions in mangrove swamps: *ibid.*, no. 4(c), pp. 102–110.

PARADICE, W. E. J. (1924) The Sir Edward Pellew Group of islands. Report with special reference to biology and physical features: Commonwealth Parliamentary Pub. 143.

——— (1925) The pinnacle- or mushroom-shaped coral growths in connection with the reefs of the Outer Barrier: Great Barrier Reef Comm. Repts., vol. 1, pp. 52–60.

RAGGATT, H. G. (1936) Geology of North-West Basin, Western Australia, with particular reference to the stratigraphy of the Permo-Carboniferous: Royal Soc. New South Wales Jour. and Proc., vol. 70, pp. 100–174.

RAINFORD, E. H. (1925) Destruction of the Whitsunday Group fringing reefs: Australian Mus. Mag., vol. 2, pp. 175–177.

RATTRAY, A. (1869) Notes on the geology of the Cape York Peninsula: Geol. Soc. London Quart. Jour., vol. 25, pp. 297–305.

REATH, J. L. (1925) Mollusca from the sub-Recent shell beds of the Lower Swan River: Royal Soc. Western Australia Jour., vol. 2, pp. 31–41.

REULING, H. R. (1934) Der Sitz der Dolomitisierung: Versuch einer neuen Auswertung der Bohr-Ergebnisse von Funafuti: Senckenberg. naturf. Gesell. Abh., no. 428, 44 pp.

RICHARDS, H. C. (1928) Scientific investigations on the Great Barrier Reef, Australia: Great Barrier Reef Comm. Repts., vol. 2, pp. vii–xvi.

——— (1931) The Grey Street Bridge, Brisbane. The rock foundations of the piers: Eng. Inst. Australia Jour., vol. 3, pp. 377–385.

——— (1938*a*) Boring operations at Heron Island, Great Barrier Reef: Great Barrier Reef Comm. Repts., vol. 4, pt. 3, pp. 135–142.

——— (1938*b*) Some problems of the Great Barrier Reef: Royal Soc. New South Wales Jour. and Proc., vol. 71, pp. 68–85.

——— (1939*a*) Recent sea-level changes in eastern Australia: Sixth Pacific Sci. Cong. Proc., pp. 853–856.

——— (1939*b*) Results of deep-boring operations on the Great Barrier Reef, Australia: *ibid.*, p. 857.

——— and HEDLEY, C. (1925) A geological reconnaissance in north Queensland: Great Barrier Reef Comm. Repts., vol. 1, pp. 1–27.

——— and HILL, D. (1942) Great Barrier Reef bores, 1926 and 1937—descriptions, analyses, and interpretations: Great Barrier Reef Comm. Repts., vol. 5, 122 pp.

SANFORD, S. (1909) The topography and geology of southern Florida: Florida State Geol. Survey 2d Ann. Rept., pp. 175–231.

SAVILLE-KENT, W. (1893) The Great Barrier Reef of Australia, London, W. H. Allen & Co., 387 pp.

——— (1897) The naturalist in Australia, London, 302 pp.

SAYLES, R. W. (1931) Bermuda during the Ice Age: Am. Acad. Arts Sci. Proc., vol. 66, pt. 2, pp. 361–465.

SCHOTT, G. (1933) Auftriebwasser an den australischen Westküsten? Ja und nein! Annalen Hydrographie und mar. Meteorologie, vol. 8–9.

——— (1935) Geographie des indischen und stillen Ozeans, Hamburg, C. Boysen, 413 pp.

SEMPER, K. (1881) The natural conditions of existence as they affect animal life, London, Internat. Sci. Ser. 31, 472 pp.

SEWELL, R. B. S. (1928) A study of recent changes of sea-level based largely on a study of coral growths in Indian and Pacific seas: Internat. Rev. hydrobiologie et hydrographie, vol. 20, pp. 89–102.

——— (1932) The coral coasts of India: Geog. Jour., vol. 79, pp. 449–465.

——— (1935) Studies on coral and coral-formations in Indian waters: Asiatic Soc. Bengalia Mem., vol. 9, pp. 461–540.

——— (1936) An account of Addu Atoll: John Murray Expedition, 1933–34, British Mus. Nat. History Sci. Repts., vol. 1, no. 3, pp. 63–93.

SHEPARD, F. P. (1948) Submarine geology, New York, Harper & Bros., 348 pp.

SKEATS, E. W. (1903) The chemical composition of limestone from upraised coral islands, with notes on their microscopical structures: Harvard Coll. Mus. Comp. Zoology Bull. 42, pp. 53–126.

——— (1905) The chemical and mineralogical evidences as to the origin of the dolomites of southern Tyrol: Geol. Soc. London Quart. Jour., vol. 61, pp. 97–141.

——— (1918*a*) The coral-reef problem and the evidence of the Funafuti borings: Am. Jour. Sci., 4th ser., vol. 45, pp. 81–90.

——— (1918*b*) The formation of dolomite and its bearing on the coral reef problem: *ibid.*, pp. 185–200.

SMITH, C. L. (1940) The Great Bahama Bank. II. calcium carbonate precipitation: Marine Research Jour., vol. 3, no. 2, pp. 171–189.

SMITH, T. H., and IREDALE, T. (1924) Evidence of a negative movement of the strand line of 400 feet in New South Wales: Royal Soc. New South Wales Jour. and Proc., vol. 58, pp. 157–168.

SOMERVILLE, J. L. (1919) Evidences of uplift in the neighbourhood of Perth: Royal Soc. Western Australia Jour., vol. 6, pp. 5–29.

SPENDER, M. A. (1930) Island-reefs of the Queensland coast: Geog. Jour., vol. 76, pp. 194–214; 273–297.

STANLEY, G. A. V. (1928*a*) The physiography of the Bowen district and of the northern isles of the Cumberland Group: Great Barrier Reef Comm. Repts., vol. 2, no. 1, pp. 1–51.

——— (1928*b*) Physiographic investigations in

Queensland with reference to the Great Barrier Reef: Am. Jour. Sci., vol. 16, pp. 45–50.

STEARNS, H. T. (1941) Shore benches on north Pacific islands: Geol. Soc. America Bull. 52, pp. 773–780.

——— (1945) Decadent coral reef on Eniwetok Island, Marshall Group: *ibid.*, vol. 56, pp. 783–788.

——— (1946) An integration of coral-reef hypotheses: Am. Jour. Sci., vol. 244, pp. 245–262.

——— and VAKSVIK, K. N. (1935) Geology and ground-water resources of the island of Oahu, Hawaii: Hawaii Div. Hydrography Bull. 1, 479 pp.

STEERS, J. A. (1929) The Queensland coast and the Great Barrier Reef: Geog. Jour., vol. 74, pp. 232–257; 341–370.

——— (1930) A geographical introduction to the biological reports: Great Barrier Reef Expedition, 1928–29, British Mus. Nat. History Sci. Repts., vol. 3, pp. 1–15.

——— (1931) Evidences of recent movements of sea-level on the Queensland coast: raised benches and the coral reef problem: Cong. géol. internat. Compte rendu, vol. 2.

——— (1937) The coral islands and associated features of the Great Barrier Reefs: Geog. Jour., vol. 89, pp. 1–28, 119–146.

——— (1938) Detailed notes on the islands surveyed and examined by the geographical expedition to the Great Barrier Reef in 1936: Great Barrier Reef Comm. Repts., vol. 4, pt. 3, pp. 51–96.

——— (1945) Coral reefs and air photography: Geog. Jour., vol. 106, pp. 232–235.

STEPHENSON, T. A.; TANDY, G.; and SPENDER, M. (1931) The structure and ecology of Low Isles and other reefs: Great Barrier Reef Expedition, 1928–29, British Mus. Nat. History Sci. Repts., vol. 3, no. 2, pp. 17–112.

STOKES, J. L. (1846) Discoveries in Australia . . . the voyage of H.M.S. *Beagle* 1837–43 . . . also a narrative of Captain Owen Stanley's visits to the islands in the Arafura Sea, London, 2 vols.

STUBBINGS, H. G. (1938) Marine sediments from islands and reefs of the Great Barrier Reef: Great Barrier Reef Comm. Repts., vol. 4, pt. 3, pp. 97–104. (Appendix to Steers, 1938.)

STUTCHBURY, S. (1854) Twelfth tri-monthly report upon the Geological and Mineralogical Survey of New South Wales: Legislative Assembly Paper, New South Wales, p. 7.

SUSSMILCH, C. A. (1923) An outline of the main topographical features of New South Wales: Pan-Pacific Sci. Cong. Proc., vol. 1, pp. 721–726.

——— (1930) Post-Pliocene tectonic movements in eastern Australia: Fourth Pacific Sci. Cong. Proc., vol. 2B, p. 657.

——— (1938) The geomorphology of eastern Queensland: Great Barrier Reef Comm. Repts., vol. 4, pt. 3, pp. 105–134.

——— and DAVID, T. W. E. (1919) Sequence, glaciation and correlation of the Carboniferous rocks of the Hunter River district, New South Wales: Royal Soc. New South Wales Jour. and Proc., vol. 53, pp. 246–338.

TAYLOR, G. (1911) Physiography of eastern Australia: Commonwealth Bur. Meteorology, Melbourne, Bull. 8, 17 pp.

TEICHERT, C. (1947) Contributions to the geology of Houtman's Abrolhos, Western Australia: Linnean Soc. New South Wales Proc., vol. 71, pp. 145–196.

——— (1950) Late Quaternary sea-level changes at Rottnest Island, Western Australia: Royal Soc. Victoria Proc., vol. 59, pp. 63–79. (In press.)

——— and FAIRBRIDGE, R. W. (1948) Some coral reefs of the Sahul Shelf: Geog. Rev., vol. 38, pp. 222–249.

TENISON-WOODS, J. E. (1878) On the extra-tropical corals of Australia: Linnean Soc. New South Wales Proc., vol. 2, pp. 292–341.

——— (1880) On a fossiliferous bed at the mouth of the Endeavour River: *ibid.*, vol. 5, pp. 187–189.

TESTER, A. C. (1948) Marine terraces of the Pacific Ocean area: 18th Internat. Geol. Cong. (Great Britain), Abstracts.

THOMSON, W. C. (1905) Upheavals and depressions in the Pacific and on the Australian coast: Queensland Geog. Jour., vol. 20, pp. 1–8.

TWENHOFEL, W. H. (1932) Treatise on sedimentation, 2d ed., Baltimore, Williams & Wilkins Co., 926 pp.

UDLUFT, H. (1929) Die Genesis der flächenhaft verbreiteten Dolomite des mitteldevonischen Massenkalkes, insbesondere des Schwelmer Kalkes der Gegend von Elberfeld-Barmen: Preuss. geol. Landesanstalt Jahrb., vol. 50, pp. 396–436.

UMBGROVE, J. H. F. (1928) De Koraalriffen in de Baai van Batavia: Dienst van den Mijnb. Wetensch. Meded. Nederlandsch Indie, no. 7, 68 pp.

——— (1929) De Koraalriffen der Duizend-Eilanden (Java-Zee): *ibid.*, no. 12, 47 pp.

——— (1930*a*) The amount of maximal lowering of sea-level in the Pleistocene: Fourth Pacific Sci. Cong. Proc., vol. 2A, pp. 105–113.

——— (1930*b*) The influence of the monsoons on the geomorphology of coral islands: *ibid.*, pp. 49–54.

——— (1930*c*) De Koraalriffen van den Spermonde Archipel, S. Celebes: Leidsche Geol. Mededeelingen, vol. 3, pp. 221–264.

——— (1931) Note on negroheads (coral boulders) in the East Indian Archipelago: K. Akad. Wetensch. Amsterdam Proc., vol. 34, pp. 485–487.

——— (1939) De Atollen en Barrière-Riffen der Togian-Eilanden: Leidsche Geol. Mededeelingen, vol. 11, pp. 132–187.

——— (1946) Origin of continental shelves: Am. Assoc. Petroleum Geologists Bull. 30, pp. 249–253.

——— (1947) Coral reefs of the East Indies: Geol. Soc. America Bull. 58, pp. 729–778.

——— and COSIJN, J. (1931) Java's Zuidkust bij Tji-Laoet-Eureun: Geol.-mijnb. genootsch. Nederland en Kolonien Verh., Geol. ser., vol. 9, pt. 2, pp. 121–134.

VAN TUYL, F. M. (1916) The origin of dolomite: Iowa Geol. Survey, vol. 25, pp. 241–422.

VAUGHAN, T. W. (1910) Contribution to the geologic history of the Floridian Plateau: Carnegie Inst. Washington Pub. 133. Dept. Marine Biology Papers, vol. 4, pp. 99–185.

——— (1914*a*) Building of the Marquesas and Tortugas atolls and a sketch of the geologic history of the Florida reef tract: *ibid.*, Pub. 182. Vol. 5, pp. 55–67.

——— (1914*b*) Preliminary remarks on the geology of the Bahamas, with special reference to the origin of the Bahamas and Floridan oolites: *ibid.*, Pub. 182. Vol. 5, pp. 47–54.

——— (1919) Corals and the formation of coral reefs: Smithsonian Inst. Ann. Rept. for 1917, pp. 189–276.

——— (1923*a*) The ecology and growth rate of corals: Pan-Pacific Sci. Cong. Proc., vol. 2, pp. 1092–1101.

——— (1923*b*) Coral reefs and submerged platforms: *ibid.*, p. 1128–1134.

——— (1940) General discussion on lists of names applied to submarine features. Report of the Committee on Criteria and Nomenclature of the major divisions of the ocean bottom: Phys. Oceanography Assoc. Sci. Pub. 8, pp. 100–118.

——— *et al.* (1918) Some shoal-water bottom samples from Murray Island, Australia, and comparisons of them with samples from Florida and the Bahamas: Carnegie Inst. Washington Pub. 213. Dept. Marine Biology Papers, vol. 9, pp. 235–287.

VERRILL, A. E. (1900) Notes on the Bermuda Islands: Am. Jour. Sci., 4th ser., vol. 9, pp. 313–340.

——— (1905) The Bermuda Islands. Pt. 4: Geology and paleontology. Pt. 5: An account of the coral reefs: Connecticut Acad. Arts and Sci. Trans., vol. 12, pp. 45–348.

VERWEY, J. (1931) Coral reef studies. 2. The depth of coral reefs in relation to their oxygen consumption and the penetration of light in the water: Treubia (Buitenzorg), vol. 13, pp. 169–198.

VOELTZKOW, A. (1901) Ueber Coccolithen und Rhabdolithen nebst Bemerkungen über den Aufbau und die Entstehung der Aldabrainseln: Senckenberg. naturf. Gesell. Abh., vol. 26, pp. 467–537.

——— (1907) Forschungen über Korallenriffe: Geog. Anzeiger, vol. 8, nos. 1–2.

WELLS, J. W.; COLE, W. S.; LILL, G.; LADD, H. S.; and TRACEY, J. I. (1948) Drilling on Bikini Atoll, Marshall Islands: 18th Internat. Geol. Cong. (Great Britain), Abstracts.

WENTWORTH, C. K., and PALMER, H. S. (1925) Eustatic bench of islands of the north Pacific: Geol. Soc. America Bull. 36, pp. 521–544.

WHARTON, W. J. L. (1883) Mangrove as a destructive agent: Nature, vol. 29, pp. 76–77.

WILKINSON, C. S. (1887) Notes on geology of New South Wales, Sydney, Dept. of Mines, pp. 49–94.

WOOD-JONES, F. (1910) Coral and atolls, London, 392 pp.

YONGE, C. M. (1930) A year on the Great Barrier Reef, London and New York, Putnam, 245 pp.

——— (1940) The biology of reef-building corals: Great Barrier Reef Expedition 1928–29, British Mus. Nat. History Sci. Repts., vol. 1, pt. 13, pp. 353–391.

ZEUNER, F. E. (1945) The Pleistocene period, its climate, chronology, and faunal succession, London, Methuen, 444 pp.

SILURIAN REEFS OF GOTLAND[1]

ASSAR HADDING
Lunds Universitets

ABSTRACT

Core drillings on Gotland in 1946 and 1947 led to a re-examination and revision of our conception of the reefs of Gotland, their structure, reef-building organisms, and appearance. A short report on the results of the investigation is submitted below. Of particular interest are the findings that calcareous algae play a considerably greater role as rock-formers than had been previously observed.

GEOGRAPHIC AND STRATIGRAPHIC DISTRIBUTION OF THE REEFS

Reefs occur in several different beds of Silurian age in Gotland. They occur in belts running southwest to northeast and were formed as barrier reefs east or southeast of a coast. The oldest are exposed in the northern part of the island, the youngest in the south.

The individual reefs vary greatly in size. Small knolls of only a few meters in diameter are fairly common in some strata; within the actual reef beds they attain a thickness of 10–20 meters and an extension of several hundreds of meters.

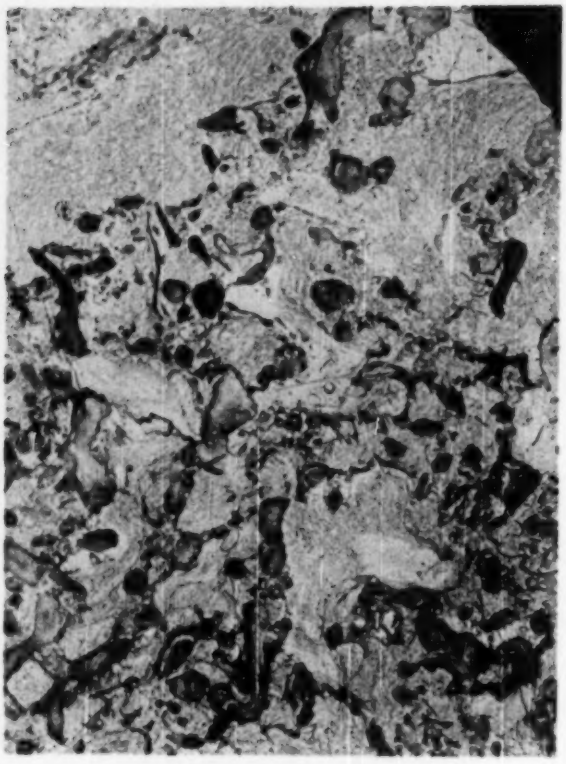
Reef-formation during any given period did not begin at one time over a large area but started from several different points. As the reef increased in height and thickness, adjacent parts joined together. Between isolated reef masses, limestone and marly shale were deposited, forming a normal continuation of the stratified rocks below the reefs, although they were somewhat modified by conditions arising from the growing reef and its organisms. Because the reef grows in both height and thickness from a comparatively narrow base, it first has the form of an inverted cone. After a short period of growth, however, many reefs spread rapidly over larger areas and assumed the form of flat lenses.

Independent of the manner of growth of the reefs, the sedimentation around them continued next to the reef formation. A prerequisite for large lateral spreading of the reefs was a suitable substratum for the reef-forming organisms. Reefs were formed on a marly substratum (as, for example, in the upper Visby marl) as well as on banks of fairly pure calcareous sand. They could not maintain their growth, however, without the support of new sediments deposited on their sides.

The reefs are entirely embedded in sediments: (1) older sediments which form the base of their oldest parts, (2) contemporaneous sediments which enclose their sides and form the substratum of the portions lying above the base, and, finally, (3) younger sediments which cover the reefs and surrounding sediments.

The reefs, or at least their central parts, rose somewhat above surrounding loose deposits during the period of formation. The sediments covering the reefs, therefore, dip away from the reefs toward the surrounding stratified layers. Compaction of surrounding beds increased the inclination unless the compaction was entirely counteracted by the upward movement which occurred around the reefs when they sank down into the substratum.

[1] Manuscript received February 14, 1950.



Reef limestone, Bungenäs, Gotland

BIOLOGIC CHARACTER AND STRUCTURE OF THE REEFS

The structure of the reefs is fairly uniform within the same stratum, but varies considerably from one bed to another. These differences are due to variations in the biologic character of the reefs, that is, to the variations in the limestone-forming and reef-forming fauna and flora.

In places where the reefs consist of ball-shaped or spheroidal stromatoporoids and coral knolls (e.g., certain forms of *Favosites* and *Heliolites*), the structure is conglomeratic and the individual reefs are commonly comparatively narrow.

In places where the reefs were formed by tabular stromatoporoids and corals, the structure of the reefs is of another character. The reefs exhibit a bedded structure and have a marked horizontal extension.

Where the reefs are formed from organisms with columnar or branched stocks, they contain a large number of fragments, together with mud deposits which have amassed and been packed around the reef-builders. The structure is either more brecciated or more massive than in the types mentioned above.

It is not difficult to determine which reef-building organisms constitute the bulk of the above-mentioned types of reefs. On Gotland, however, there are also typical reefs in which no actual reef-builders are obvious. Scattered stromatoporoids and corals occur, but they do not form a continuous framework. The interstitial matter, consisting of fragments and fragment-mixed mud, does not entirely fill up the cavities between these calcareous bodies. The reefs are composed mainly of a dense, compact mass which surrounds corals, stromatoporoids, and other calcareous organisms, as well as the accumulation of fragments (pl. 1). This dense mass, which the author has previously interpreted as consolidated, pure calcareous mud, is, however, a product of calcareous algae. The faint "stratification" which is discernible in the dense mass consists of irregular layers which follow the varying form of the substratum, both horizontally and vertically. There was a continuous struggle between algae and corals, and the structure of the latter has decomposed into a thin, irregular network. Reefs of this type—algal-calcareous reefs—have a compact structure and a comparatively flat form.

STRATIFIED ROCKS SURROUNDING THE REEFS

The strata which surround the reefs on Gotland consist of stratified limestone in part interbedded with marly shales. In the peripheral areas of a reef, the stratified rock and the reef may interfinger because of the alternate expansion and reduction of the reef during its period of growth.

The substratum of the reefs was deposited as normal bedded limestone or as marl with slight dip or none at all. It is now found dipping in under the reefs as a result of the sinking of the bulk of the reef. Only in this manner has the rock of the substratum been influenced by the reefs.

On the other hand, the limestone which was deposited on the sides of the

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PLATE 1

Reef limestone with slender Coenites branches (dark) surrounded by a mass of calcareous algae which form the actual framework of the reef. Bungenäs, Gotland. Natural size.

reefs and which was formed at the same time as the latter consists more or less of fragments of the reef structure washed away from the reef or of organisms living on the reefs. The composition of these sediments and their variation away from the reefs have been described by the author in a previous publication (1941). An investigation carried out during recent years has also yielded such a clear picture of the distribution and general character of the sediments within the area adjacent to some of the reefs that it is worthy of mention.

In and around two reefs at Smöjen in the northeastern part of Gotland, some 70 core drillings were made during 1946 and 1947. The strata belong to the Slite stage and consist of the following parts: upper Smöjen limestone (*Megalomus* limestone); middle Smöjen limestone (Smöjen reef horizon); lower Smöjen limestone (Smöjen dark-colored limestone); and Slite marlstone.

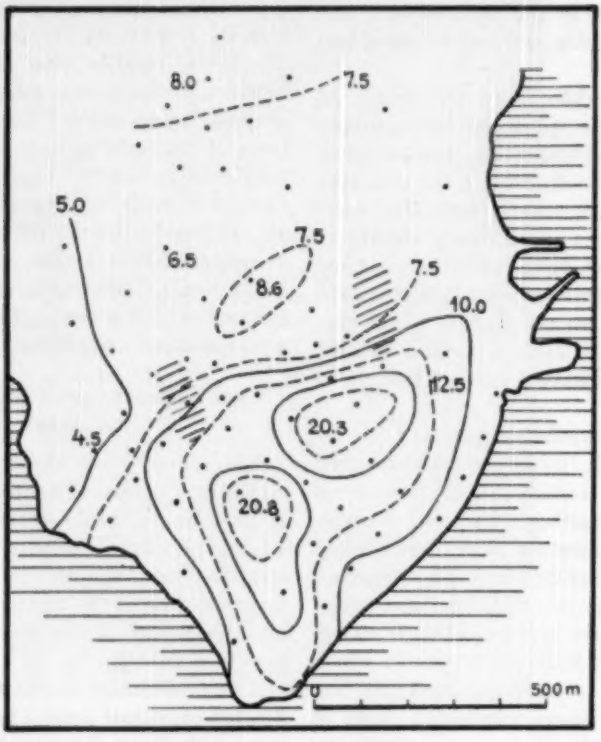


FIG. 1.—The Smöjen district. Sketch showing the thickness and distribution of the Smöjen limestone reefs and stratified limestone with marly part (hatched areas). Dots represent core drillings. Elevations in meters.

The accompanying sketches show the horizontal distribution of the sediments around the reefs (figs. 1–2) and, in two sections (fig. 3), the vertical distribution and location of the reefs and stratified sediments in relation to the sea level. The

following points should be emphasized: (1) the secondary influence of the reefs on the primary horizontal position of the substratum: strongest depression under the thickest part of the reef and the stratified limestone; and (2) the primary influence of the reefs on sedimentation: strong accumulation of fragments and formation of large banks of limestone close to the reefs; deposition of marl on the lee side of the reefs.

The movement of the water around the reefs (surfs, tidal currents, etc.) was stopped in some places and had free play in others. The result has been the deposition of muddy sediments alongside pure limestones. While the reefs were still young and only a few meters thick, marly layers were deposited on both the lee and windward sides. Only in the channel between them was the movement of the water too strong to allow of any deposition of mud. When the reefs were fully developed, only a limited area on the lee side was suited for mud deposits. Clean-swept fragments formed banks of pure limestone on each side of the reef, broken only by a marly streak on their northwest side facing the coast.

Coarse fossil fragments and large fragments of the reef formation might be expected nearest to the reefs and fine-

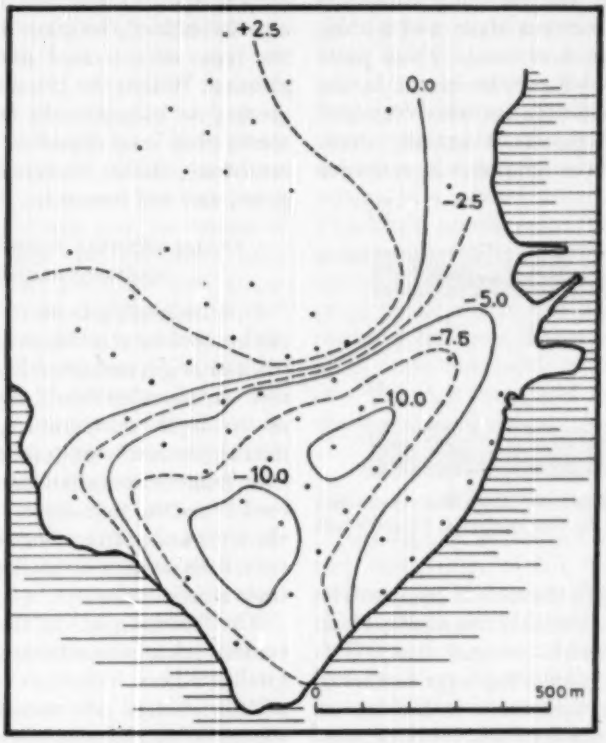


FIG. 2.—The Smöjen district. Sketch showing the level of the foot wall of the Smöjen limestone. Sea level is 0.0. Elevations in meters.

grained material to be more widely distributed. This is generally the case, but at the Smöjen reefs it is difficult to demonstrate such a distribution. The reason is that these reefs contain no large resistant calcareous bodies like stromatoporoids or powerful coral stocks. As mentioned above, the reefs are composed mainly of calcareous algae and a thin, irregular network of corals. Those parts of the reefs which were broken off by the surf before consolidation were crumbled into small fragments. Unusually even-grained mud-free limestone in extensive

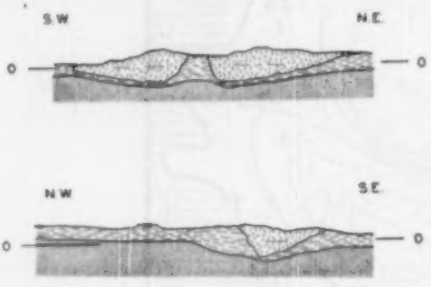


FIG. 3.—Sections from Smöjen. Reefs in stratified limestone above the Slite marlstone. Height : length = 2 : 1.

banks surrounds the reefs at Smöjen. The accumulation of crinoid fragments, which is so characteristic around many Gotlandian reefs, producing large masses of pure crinoid limestone, is lacking at Smöjen. The Smöjen reefs, it is true, contain crinoid fragments, but only in moderate quantity.

REEF-FORMATION DURING ELEVATION OR SUBSIDENCE OF LAND

In some cases, it is possible to determine whether a reef has been formed in connection with a change in level. The oldest large reefs on Gotland, the Högklint group, were formed during a period of regression. This is proved by the ascending succession of strata, from marly shales through marly and pure stratified limestone of the reef zone to pronounced algal limestone and intraformational conglomerate. The reefs in the Slite group (the above-mentioned reefs at Smöjen) were also formed during a period of land elevation. In other cases the reefs were formed during a transgression. Within the Hamra and Sundre groups, for example, the following sediments have been deposited successively: sandstone, oölitic limestone, algal limestone, and reef limestone.

ENVIRONMENTAL CONDITIONS OF REEF-FORMATION

It is impossible to learn in detail the environmental conditions existing at the place of origin and growth of a Paleozoic reef. An investigation of such conditions as the depth, temperature, and muddiness of the water can be carried out, however. Some information can also be gathered from the petrological character of the reefs and surrounding sediments and from their content of fossils and mud and their stratification.

The Silurian reefs of Gotland always contain calcareous algae and must therefore have been formed in comparatively shallow waters. As mentioned above, they may have developed during a period of changes of level, even during an elevation of land, and therefore the various parts may have been formed at various depths. The author has calculated the depth to be between 5 and 50 meters (1941).

The temperature of the water around the reefs cannot be estimated from the corals, bryozoans, or other organisms which are present. It may be assumed, however, that the intense growth of lime-

forming fauna and flora must have taken place in a comparatively warm sea. It is hardly possible, however, to state the temperatures within which this development has occurred or to consider the variations in the lime-deposition as evidence of variations in temperature. Other environmental changes may have played a greater part in their development.

Many of the reefs on Gotland are embedded in marly strata, and the reefs themselves are then mingled with mud. The organisms which have built these reefs did not require a mud-free water to thrive. They have, on the contrary, served as mud-traps and, by means of accumulated mud and fragments, have obtained support for their own structure. Several of them appear also in marl and have succeeded in creating a firm base for a reef formation in these soft sediments also. *Halysites* and *Favosites* are commonly found in such cases, as well as dome-shaped stromatoporoids.

The portions of the reef that are surrounded by contemporaneous, pure, fragmentary limestone contain very little mud or none whatsoever. It is not a coincidence that the organisms which have built these reefs are of a somewhat different kind than those formerly mentioned. The remarkable role played by calcareous algae in the building of the pure reef limestones on Gotland should be noted. These algae are of an essentially different type than those found in other reefs, for example, the Ordovician-Silurian reefs in middle Sweden (Dalecarlia). They will therefore be discussed more in detail in the following paragraphs.

Although the reefs of Gotland are marine, as shown by the fauna, and are shallow-water formations, as shown by the flora, they were not formed within a coastal area with a heavy surf. Waves and currents have not been able to break to pieces the commonly remarkably brittle skeleton of the reef. The common fine-grained sediments deposited close to the reefs also indicate formation in comparatively quiet water. Smooth stromatoporoids and compact resistant corals, which characterize certain reefs or parts of reefs, indicate, however, that a certain selection may have taken place among the reef-builders, conditioned by a greater motion of the water during a certain period and in certain places resulting in a greater strain on the reef. This strain appears also from the larger accumulation of coarser fragments from the reef, which locally form conglomerate banks within the limestones surrounding the reefs. These conglomerates occur most commonly with those parts of a reef that have been formed in the shallowest water.

THE CALCAREOUS ALGAE IN THE REEFS OF GOTLAND

Although a systematic investigation and description of all reef-builders—stromatoporoids, corals, bryozoans, and calcareous algae—still remains to be carried out, they are well known as rock-formers. Very little is known of the calcareous algae, however, and they have hitherto been underestimated. A few remarks will therefore be made on their occurrence and different forms.

Rothpletz (1908, 1913) has described the algae of Gotland which appear in the form of rounded or irregular balls: *Sphaerocodium*, *Spongiostroma* (interpreted as a hydrozoan by Rothpletz), and *Solenopora*. The algae of the first two groups are built concentrically around a nucleus, usually a shell, coral fragment, etc. The *Sphaerocodium* balls

have winding and, in some cases, branching algal fibers. *Spongiostroma* have no typical structure, but the primary porous mass has an irregular concentric or bulb-shaped form.[2] Both *Sphaerocodium* and *Spongiostroma* occur in banks in the form of accumulated balls within different parts of the Gotlandian series of strata. They always indicate sedimentation in shallow water.

The *Solenopora* knolls have a radial structure fairly similar to that of *Lithothamnium*. Unlike the *Sphaerocodium* and *Spongiostroma* balls, the *Solenopora* lumps are a normal element in the reefs of Gotland. They are easily observed and identified because of their structure and characteristic porcelaneous luster.

In addition to these well-known algae, there are also others within the reef limestone which are of greater importance for the actual reef-formation. Mats of calcareous algae with distinct fibrinous structure occur very commonly, connecting the loose fragments in a reef.[3] Large continuous parts of identical or similar structure may form considerable portions of a reef.

In the massive reefs (within the Slite stage), rock-forming calcareous algae are seen. They are irregularly built, with indistinct spongy structure. They not only connect loose fragments but also form the actual firm bulk within large parts of a reef. Stages of growth can be observed in some portions of these algal masses, in some cases parallel but never quite regular. In others, the growth has taken place more freely and irregularly. The mass has grown upward and sideward, enveloping corals, stromatoporoids, and fragments. The characteristic blurred appearance of this algal structure most closely resembles that of *Spongiostroma*. A closer examination would probably reveal that more than one algal species has taken part in the formation of the limestone.

The *Solenopora* algae appear within different parts of the reefs on Gotland, in some cases in isolated lumps or in some cases in greater numbers as irregularly shaped, many-branched masses. As reef-builders, they are less significant than the above-mentioned types of algae.

[2] The *Girvanella* balls alternate with *Sphaerocodium* and *Spongiostroma* lime.

[3] These types of algae have been previously mentioned by the author under the denomination of *Pilotrix* (1941).

SUMMARY

Reefs occur within various stages of the series of strata of Gotland, from the Högklint group (lower Wenlockian) to the Sundre group (upper Ludlovian).

The reefs appear as large or small masses, commonly with a thickness of 20 meters and a lateral extent of several hundred meters. They are arranged in belts running northeast-southwest. The oldest reefs are exposed in the northern parts of the island, the youngest in the southern part.

The structure of the reefs varies with their biologic character, particularly with the type of reef-builders.

The reefs have been formed in a warm sea in shallow water, commonly in connection with changes in sea level, positive or negative.

In some cases, the reefs or parts of them have been formed in muddy water; in other cases, in clear water.

Together with stromatoporoids and corals, calcareous algae play a major role as reef-builders. Various types are rock-forming.

Through their influence on currents

and wave motion the reefs have affected the sedimentation in their vicinity, especially the distribution of coarse and fine material.

The stratified limestones and marls, deposited on the flanks of the reefs, have grown at almost the same rate as the reefs. During the development their upper surface has lain almost on a level with (only slightly below) that of the reef.

REFERENCES CITED

HADDING, ASSAR (1941) The pre-Quaternary sedimentary rocks of Sweden. VI. Reef limestones: Lunds univ. årsskrift, new ser., pt. 2, vol. 37, no. 10. Kungl. Fysiogr. Sällsk. Handl., new ser., vol. 52, no. 10.

ROTHPLETZ, A. (1908) Über Algen und Hydrozoen im Silur von Gotland und Ösel: K. svenska vetensk. akad. Handl., vol. 43, no. 5.

——— (1913) Über die Kalkalgen, Spongiostromen und einige andere Fossilien aus dem Obersilur Gottlands: Sveriges geol. undersökning, ser. Ca, no. 10.

ORGANIC GROWTH AND SEDIMENTATION ON AN ATOLL[1]

H. S. LADD, J. I. TRACEY, JR., J. W. WELLS, AND K. O. EMERY
United States Geological Survey

ABSTRACT

An atoll is a composite reef made up of a number of distinctive reef types—windward reefs, leeward reefs, and lagoon reefs. Most of these are zoned relative to prevailing winds and currents. Organic growth, erosion, and deposition of sediment influence each of these zones, but in each the balance of forces operating at the present time may be roughly appraised. Some reef zones appear to be making headway against the sea, others appear to be essentially in a state of equilibrium, while still others are being eroded. Organic growth is the source of all the materials that make up an atoll. Some of the skeletons of the reef-building organisms remain in position of growth after death, but a much larger proportion are broken up by physical and organic agencies to form sediments that are deposited on the seaward slopes or in the lagoon. So long as organisms live on an atoll, it continues to grow, but it grows mainly by the accumulation of clastic sediments.

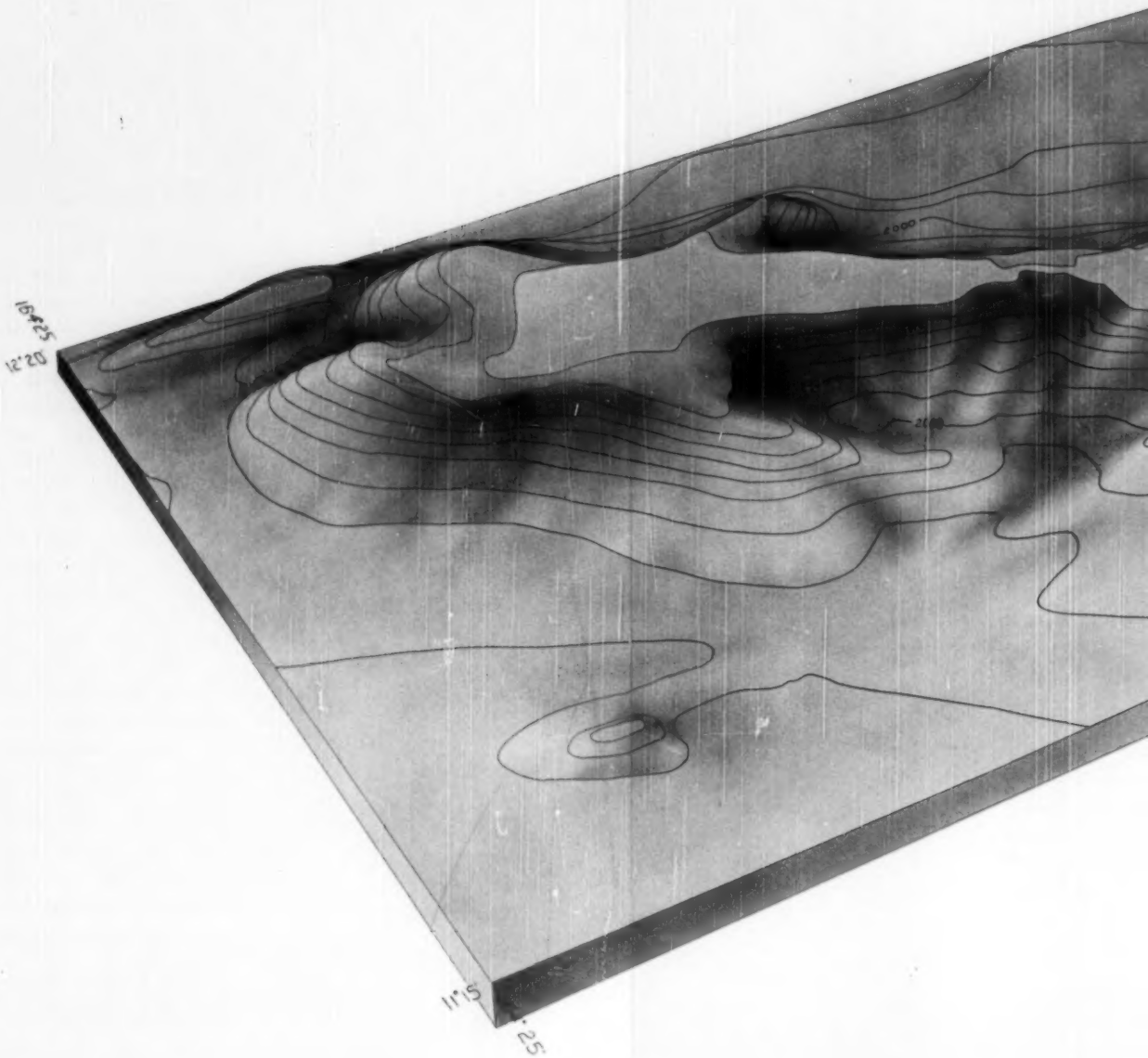
INTRODUCTION

Coral reefs are generally regarded as fairly simple structures, composed mainly of corals or of algae, that may be assigned to one of three classes—fringing, barrier, or atoll. Actually, every reef is a complicated structure even when judged from the characteristics of its surface alone. Most reefs exhibit well-developed zonation, the zones tending to parallel the reef front. The character and width of these zones are controlled by the prevailing winds and currents and by many other ecological factors. These same ecologic factors are responsible for lateral changes. Both types of changes—transverse and longitudinal—may be recognized on fringing reefs and barriers that skirt continents or large islands, but a complete pattern is most clearly seen on atolls and on the barrier reefs that completely encircle small islands. The purpose of this paper is to point out the several kinds of reef areas that may be recognized on an atoll and to describe and evaluate the zones of organic growth, the zones of erosion, and the zones of deposition that characterize each type. Such description and evaluation lead to a recognition of the balance of forces acting on the several parts of an atoll during normal times and during damaging storms. Parts of the reef are growing, some laterally, others vertically; some parts appear to be in a state of equilibrium under present-day conditions, others are adjusted to ordinary conditions but undergo severe losses during periods of heavy storms.

On atolls marine organisms are the sole builders of the existing reef. Part of the reef, especially the marginal zone, is composed dominantly of organisms that have grown in place, but the great bulk of the atoll is composed of sediments, made up of more or less fragmentary skeletons and shells.

Atolls, like other types of coral islands, are not all exactly alike. Even those in the same island group may show some striking differences. In describing the reef zones of Bikini, we do not intend to predict that identical or even very similar zones will be recognized in all other parts of the coral seas. We do know, however, that the major features of Bikini are quite similar to those of its near

[1] Published by permission of the Director, U.S. Geological Survey. Based on geological work done at Bikini with the support of the Military Intelligence Division, Office of the Chief of Engineers and the Bureau of Ships. Manuscript received January 3, 1950.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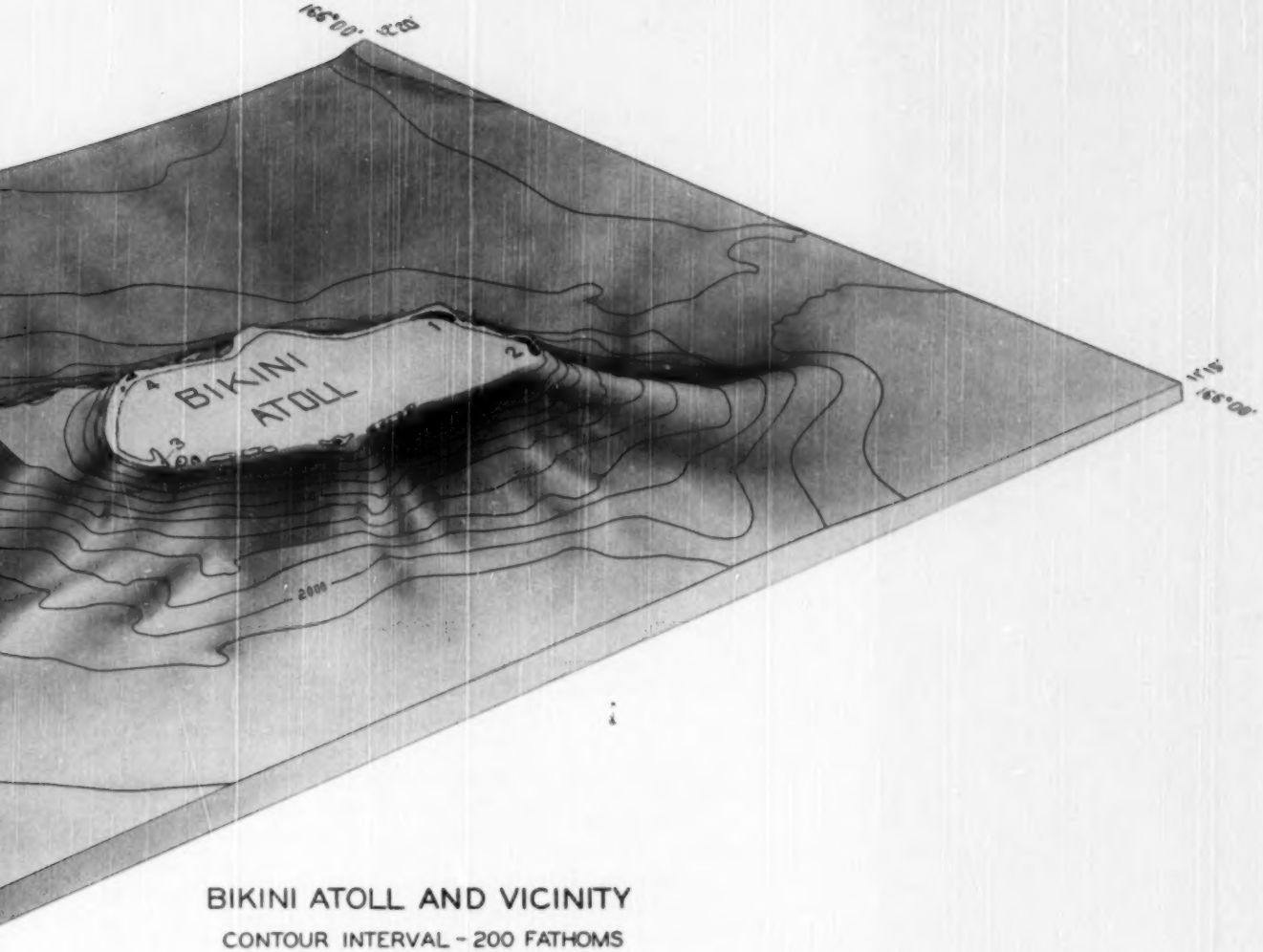


BIKINI ATOLL AND VICINITY

CONTOUR INTERVAL - 200 FATHOMS
VERTICAL EXAGGERATION - 2

1 - BIKINI ISLAND
2 - ENYU ISLAND
3 - BOKORORYURU ISLAND
4 - NAMU ISLAND

neighbors in the Marshall Islands, several of which were studied somewhat less intensively than Bikini was. The control exercised by wind and wave and temperature is great, and, where these forces are essentially the same, the reefs produced have much in common.

WINDWARD REEFS

At Bikini (fig. 1) the prevailing winds are the trades that blow from the east and northeast during most of the year. Periodically during the summer months (June–August) the trades are overcome by southerly winds, and at long intervals severe storms may arrive from that direction. The effects of this unequal distribution of winds and of the waves produced by them are clearly reflected in the form of the reef and in the distribution of some of the structures associated with it (pl. 1). The windward reefs of the north and east differ markedly from the leeward reefs of the west; reefs on the south side that are subjected to heavy storm attack fall into still a third class. The sides of the atoll that are directly exposed either to the prevailing trades or to the southerly winds are bordered by a shallow terrace that slopes very gently seaward, from a depth of 15–25 feet at the reef edge to a depth of about 50 feet at the outer edge. Where well developed, the terrace is rather flat for a distance of several hundred yards from the reef edge; in other places its slope is 10°–15°. The living reef rises from this terrace. The upward slope from the terrace to low-tide level is comparatively steep (pl. 3, *A*) and forms a distinctive zone, the remainder of the reef surface being comparatively flat. On this flat other zones parallel the reef edge. A particularly good example of reef zonation is shown in

FIG. 1.—Location of Bikini Atoll

PLATE 1

Perspective diagram of Bikini and vicinity showing seamount (guyot) extending to west. (Outline and contours from K. O. Emery; airbrush shading by Ossie Goodloe.)

plate 2, where the marginal zone and reef flat may be divided into the following zones:

A. Seaward slope to terrace (visible at upper left of pl. 2)
B. *Lithothamnion* ridge
C. Coral-algal zone
D. Outer *Heliopora* zone
E. Main reef flat
F. Inner *Heliopora* zone
G. Beach

Over large sections of the reef living organisms occur in profusion and, considered as a whole, the reef appears to be flourishing. The marginal zones appear to be growing seaward, and some of the other zones are growing upward. However, very large sections of the reef flat appear to be static, with no appreciable changes due to growth, erosion, or deposition; one small area appears to be undergoing active erosion. Each of the zones is briefly described below, and an appraisal of the present trend is given.

SEAWARD SLOPE TO TERRACE

In the area shown in plate 2 this submarine slope measures approximately 150 feet in width, sloping at an angle averaging about 15° from low-tide level to the surface of the terrace. At fairly regular intervals along the slope there are roughly straight grooves normal to the reef front. These grooves are from 5 to 10 feet in width, up to 25 feet deep at their inner ends, and their rock floors, conformable with the terrace, are covered in some places by a veneer of sand, gravel, or boulders. The grooves are separated by flattened spurs, 25 feet or more in width.

The spurs were not sampled, but the authors examined them by swimming with a face mask. They are composed mainly of living algae, but there is no evidence to indicate that the algae are roofing over or otherwise encroaching significantly upon the grooves. The seaward ends of the spurs are bluntly rounded, and they are believed to be growing outward and upward, slowly extending themselves over the terrace.

Munk and Sargent (1948) have indicated that the Bikini reefs are dynamically adjusted to prevailing ocean waves and have shown that the distribution of the marginal grooves can be correlated with the mean annual distribution of wave power around the atoll. The figures they give are impressive: they state that against the windward side of Bikini alone the waves dissipate 500,000 horsepower, one-fourth the power generated at Hoover Dam. This force, however, is not very effective in eroding the reef. Under weather conditions prevailing at times

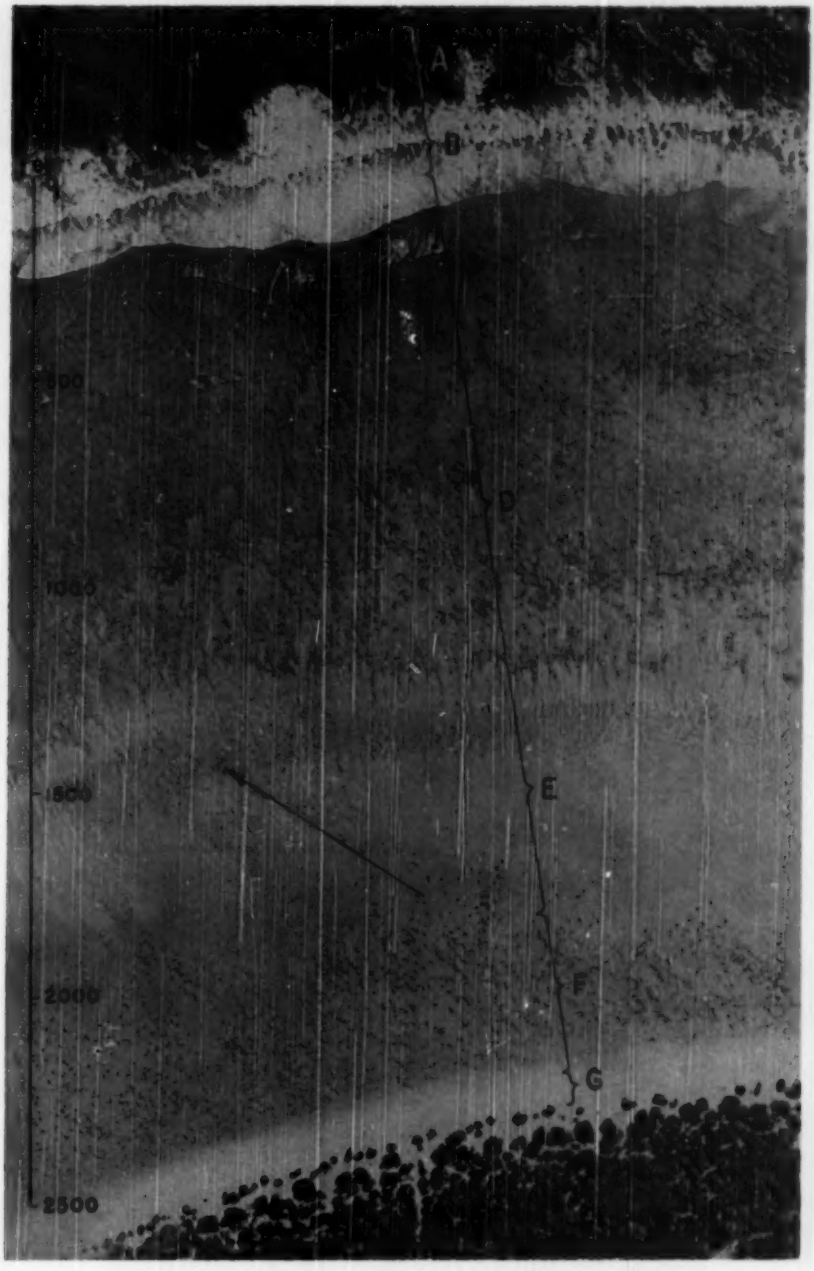
PLATE 2

Zonation on the windward reef off Bikini Island. *A (to left of bracket)*, grooves and spurs below low tide; *B*, *Lithothamnion* ridge, largely hidden by surf; *C*, coral-algal zone; *D*, outer *Heliopora* zone; *E*, main reef flat with old reef line; *F*, inner *Heliopora* zone; *G*, beach. (Aerial photograph by U.S. Navy.)

PLATE 3

A, Algal spurs and grooves bordering a windward reef. The spurs rise from a submarine terrace about 20 feet below the surface. The reef flat is to the right. The spurs are about 150 feet long, the grooves between them from 2 feet to 10 feet wide. This reef lies northwest of Bikini Island, faces north, and receives the trade swells that are refracted around the convex arc of reef off Bikini Island. This arc partly protects the reef in the area shown in upper left of photograph, and fewer grooves are developed there. (Air photograph from about 500 feet altitude by R. Dana Russell.)

B, Underwater view in surge channel near seaward edge of windward reef. Depth of channel about 12 feet. Note growth of coral on bottom and left wall. (Photograph by R. Dana Russell.)



Windward reef zonation, Bikini Island

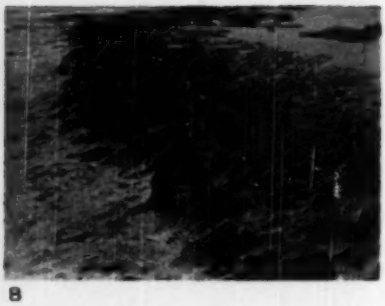
Windward reefs: spurs, grooves and surge channel

A

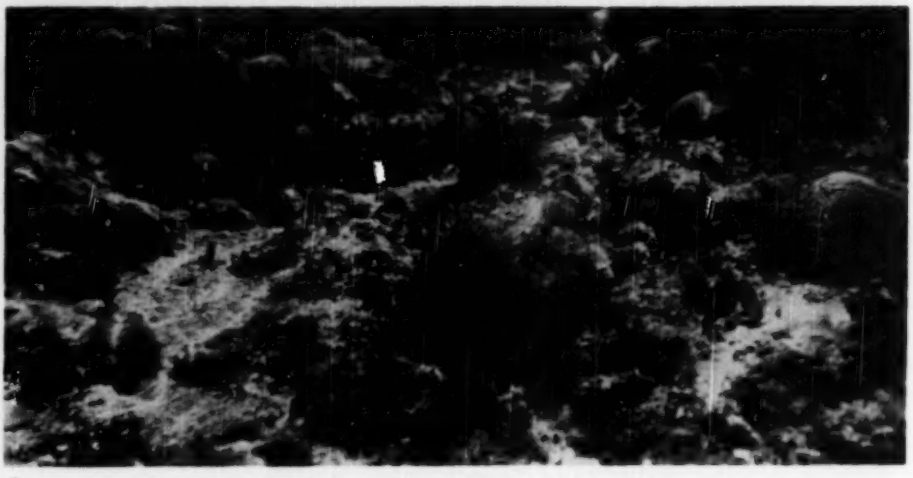
B

Lithothamnion ridge and colonies

A



B



C

Microatolls, beach-rock belt and coral-algal zone

of observation, such clastic material as is present on the floors of the grooves is not churned up and ground against the reef with each rhythmical wave advance, although on occasion the waters off the reef edge were found to be slightly murky. Presumably, the clastic material passes seaward down the grooves and across the terrace. Probably there is mechanical abrasion during periods of exceptionally heavy weather, but this does not seem adequate to explain the grooves as erosional features. Kuenen (1933) and Hanzawa (1942) have discussed these grooves ("channels," "toothed edges"). They agree that growth rather than erosion or solution is primarily responsible. Kuenen believes that the toothed edges of the East Indian reefs have been eroded since their formation, citing those described by David and Sweet (1904) from Funafuti as living examples in the open Pacific. The presence of grooves is apparently controlled by exposure to waves, and their length is influenced by the slope of the reef and by other factors. However, the most striking features of the zone where the grooves occur are the spurs, and these appear to be growth forms. The grooves, and the surge channels into which some of them lead, form a most effective baffle that dissipates the destructive energy of the waves and at the same time brings a constant supply of fresh water with oxygen, food, and nutrient salts to a maximum surface area.

Lithothamnion RIDGE

At the seaward edge of the reef a rich growth of calcareous algae forms either a gently convex rim or a cuesta-like structure (pl. 4, *A*) whose steeper side faces the sea. The convex rim is formed chiefly along the stretches of reef that lie between islands, the cuesta-type ridge where the reef fringes islands, especially where the islands lie inside arcs of reef that are convex to the sea. The rim type rises only 6 inches to 1 foot above the main reef flat, but the cuesta type may rise as much as 3½ feet. The upper limit of each appears to be determined by the height to which the waves can wash

PLATE 4

A, Gentle backslope of cuesta-like *Lithothamnion* ridge exposed at low tide. Note surge channel in right middle ground that has brought in foam-covered water near center of picture. Jagged masses projecting above surface on left are remnants of an older *Lithothamnion* ridge. The organisms forming the older structure are like those of the existing ridge, but none are alive; the older ridge is now being destroyed by solution, wave erosion, and boring organisms. It may have existed as a marginal ridge in the past, when the sea stood 6 feet higher. (Photograph by J. I. Tracey, Jr.)

B, Growth of globular colonies of *Lithothamnion* along extension of surge channel into main reef flat. This type of alga requires constant circulation of water and attains its best development on and near the seaward face of the *Lithothamnion* ridge. (Photograph by U.S. Navy.)

PLATE 5

A, Typical microatolls in the inner *Heliopora* zone (Pl. 2, zone *F*). Dark borders are living *Heliopora* except for the near side of microatoll at upper left, which is composed of *Porites*. Flat central areas are veneered with calcareous algae. (Photograph by J. I. Tracey, Jr.)

B, Solution basins and miniature karst topography developed on seaward part of beach rock belt. (Photography by K. O. Emery.)

C, Underwater view in shallow pool in coral-algal zone near Bikini Island. Bottom of pool is only 2 feet lower than reef flat and is partly covered with coarse calcareous sand. Growing on bottom are colonies of *Favia* and *Astreopora;* tufts of green algae are common. Much of the submerged reef flat seen on the lee reefs greatly resemble the conditions shown in this windward pool. Base of photograph represents about 4 feet. (Photograph by J. I. Tracey, Jr.)

regularly at low tide, because algae of the types growing on the reef edge cannot survive desiccation—they require, in fact, considerable circulation of agitated water.

Some of the grooves of the seaward slope are continued into or through the *Lithothamnion* ridge as surge channels (pl. 3, *B*) for distances of 50 to as much as 300 feet. Pavement-type algae cover the back slope of the ridge and form thickened rims along the edges of the channels to which each wave brings a fresh surge of water (pl. 4, *B*). The algae, by growing laterally at the surface, eventually roof over the landward parts of the surge channel, thus adding new reef surface and creating caverns beneath the reef. Before the roofing-over process is completed, blowholes may be developed, as described by Tracey, Ladd, and Hoffmeister (1948). From some of these the water issues from geyser-like algal craters, in other areas from a series of holes along an old fissure; in still other places hundreds of holes may form a honeycomb of openings. At low tide the geyser-type holes spout water violently and noisily with each incoming wave; from the others, in most cases, the water wells quietly, the mounds of water merging to form a thin sheet that washes landward over the flat.

The vigorous growths of algae along the sides of the surge channels and in areas of blowholes indicate clearly that the algae in those areas are making headway against the sea. The passages and caverns below the surface support a rich growth of corals which probably in time will fill much of the space there. The algal pavement that forms the gentle back slope appears to be fairly well stabilized in most places. Surge channels are being roofed over in many places, thus extending the area of the algal pavement. In other areas it is riddled by echinoid borings. The edges of the older borings are covered, however, by new algal growth, and it is doubtful whether the echinoids can do permanent damage to the algal ridge. The buttresses that face the sea and receive a heavier pounding than any other part of the reef are completely covered by living calcareous algae; parts of them are covered by the smooth botryoidal crusts of pavement algae, and all channels and depressions are lined with globular masses of branching *Lithothamnion*. The thick branches of these clusters form porous masses capable of absorbing the force of heavy waves without breakage.

CORAL-ALGAL ZONE

On the inner side of the marginal zone there is a belt of rich coral growth nearly 400 feet wide (*Acropora cuneata* and *digitifera* zones, fig. 2). In much of this area true reef corals cover 50 per cent of the area or more. The remainder of the zone is mostly covered with pavement and nodular types of pink *Lithothamnion*. The pavement algae surround heads of living coral and in some places seem to be encroaching upon them; in other places the roles are reversed, and corals appear to be encroaching upon the algae.

The coral-algal zone appears to be fairly well adjusted to existing conditions. The growth forms of the coral colonies—encrusting, low, and massive or with short, stubby branches—do not extend above the reach of waves at low-tide level. It is clear that they cannot rise higher without drying at times of low water.

OUTER *Heliopora* ZONE

In this area the general surface of the reef flat is distinctly lower than the zones to seaward and is covered by 1–4 feet of water at low tide. From the floor large

subcircular masses, 3–25 feet or more in diameter, consisting largely of the blue alcyonarian, *Heliopora*, rise close to low-tide level. In the outer part of this zone the microatolls are formed mainly by the scleractinian coral, *A. palifera*. This zone is a part of the belt of microatolls shown in figure 2. In these structures, as in true atolls, there is a concentration of living forms around the periphery, although live colonies, particularly *Heliopora* and algae, may be growing sporadically all over the structure. The *Acropora*, which apparently is not quite so hardy a form as the *Heliopora*, is concentrated around the edge of the microatoll, and most of the colonies are at a somewhat lower level than the Helioporas; at low tide the tips of many *Heliopora* colonies break water.

The upward growth of the microatolls is definitely limited by low-tide level, but the growth around the rim of each structure is rich, and the masses appear to be expanding laterally in all directions. As they coalesce, they form a new reef surface—not as firm a surface as the pavement from which they grow but one that may become so eventually by continued organic growth and silting. Many of the dead Helioporas in the centers of the microatolls are covered by a film of sand very rich in Foraminifera.

MAIN REEF FLAT

Between the outer and inner *Heliopora* zones lies the main reef flat, a fairly level rock surface that is divided by what appears to be the margin of an older reef—one that may have flourished at a time before the part of the reef now lying to seaward had grown up to low-tide level and had thus effectively reduced circulation. At the present time this old reef line is a hummocky surface of limestone cut by wide transverse gaps and narrow longitudinal fissures. It is practically bare of organic growth, and parts of it that dry at low tide appear to be undergoing erosion by solution. The surface of the main flat on either side appears to be stabilized under present conditions. The main flat supports a variety of organisms, but there is no great development of reef-building corals or algae. Much of its surface is veneered with a mat of smaller Foraminifera held together by a fibrous alga, but the Foraminifera are not building up the reef surface. As they die, their shells are carried landward by waves, and they form the main constituent of the sands of the beaches.

INNER *Heliopora* ZONE

In this zone the microatolls (pl. 5, *A*) are more widely scattered and in most places are smaller than in the outer zone. *Heliopora* is the dominant form, but, nearer shore, colonies of *Porites lutea* become increasingly numerous and in some structures almost completely replace the *Heliopora* (fig. 2). Nodules of living coralline algae 1–2 inches in diameter are abundant. If existing conditions continue, it seems likely that the microatolls will continue to spread and will eventually cover most if not all the intervening pool areas.

BEACH

The beach itself, like the island it fringes, appears to be fairly well stabilized and adjusted to existing conditions. There is a strip of coarse sand and gravel at the edge of the inner *Heliopora* zone, above which lie the cream-colored foraminiferal sands of the beach proper. There is a systematic migration of dead foraminiferal shells around the ends of Bikini Island to the beaches of its lagoon side. Many of the shells of the "beach Foraminifera" find their final resting place in the lagoon. A few are found scat-

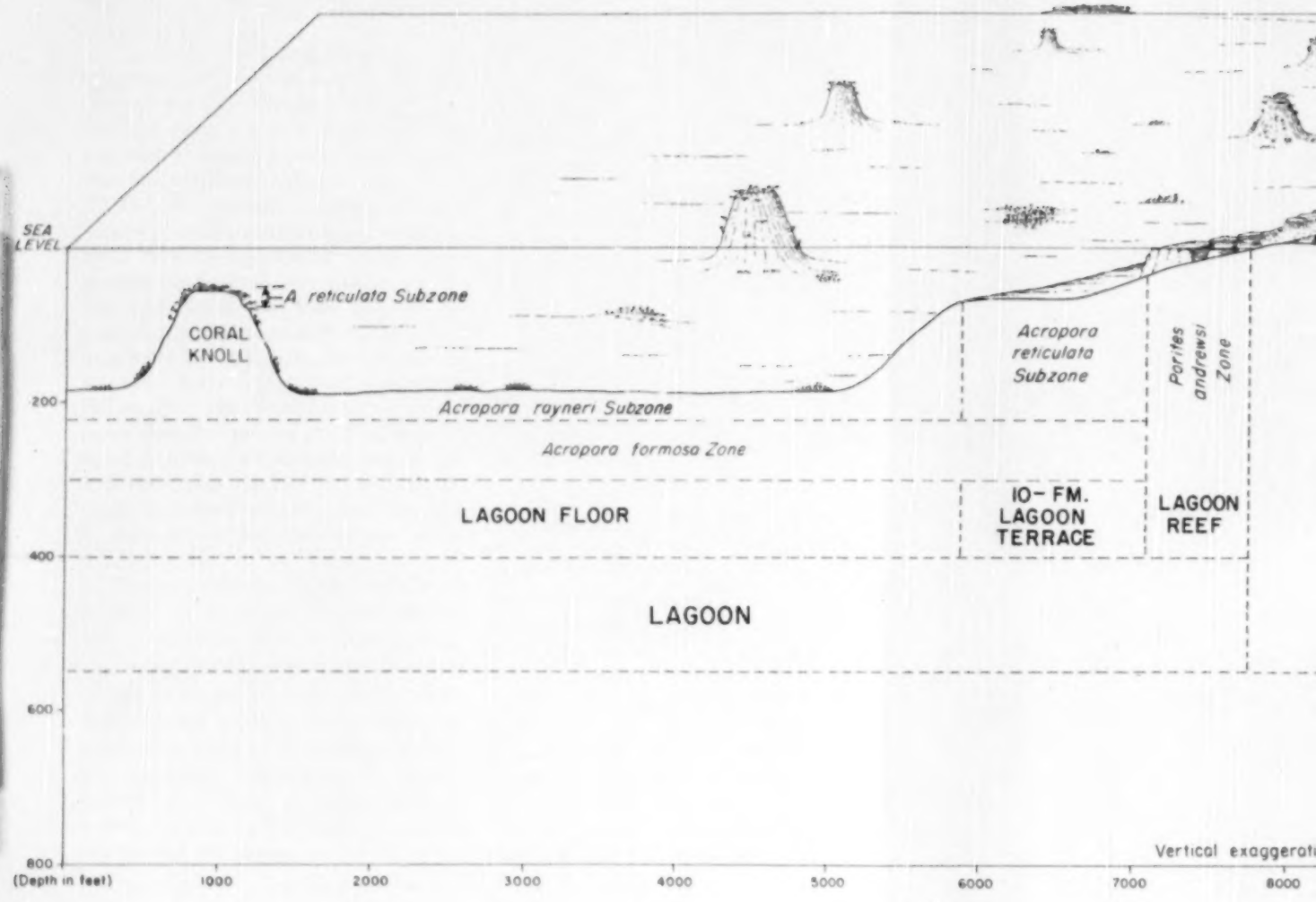
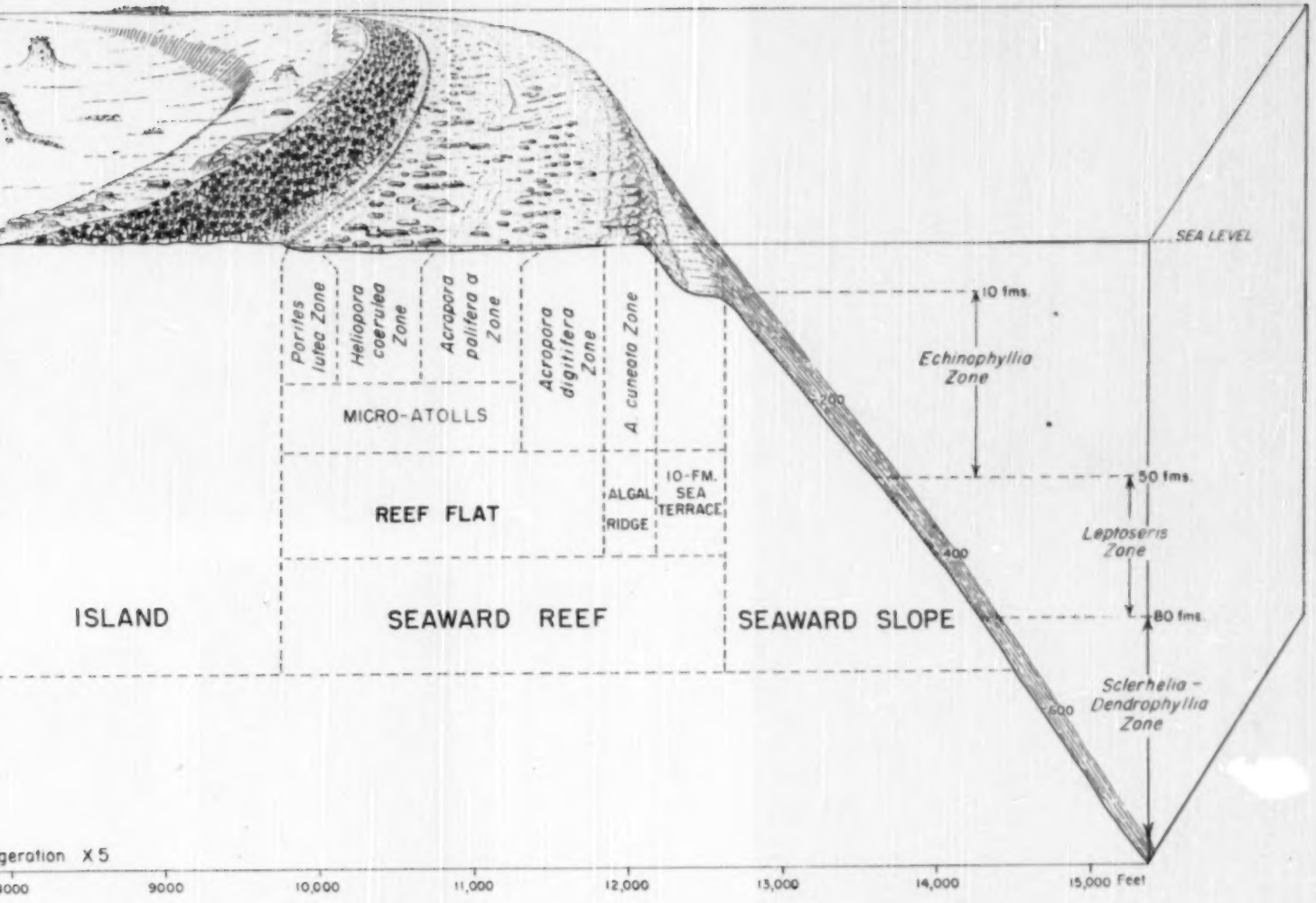


FIG. 2.—Block diagram with zonal analysis of coral fauna of Bikini Atoll. As shown in this section, the microatolls near the island



island are on the average larger than those to seaward, but in many places the reverse is the case. (Drawn by J. W. Wells.)

tered through the sands of the lagoon floor several miles from the reef, but heavy concentrations are found only near the margin.

A zone of beach rock (sandstone or conglomerate) may be formed between the reef flat and the beach. The lithified material rises 1–3 feet above low tide and may extend as a rock floor beneath the unconsolidated sands of the beach. All but the coarsest conglomerates are bedded, the layers in most places dipping away from the island at low angles comparable to those of the present beach. The rock has a fresh look, and the organic fragments retain such color as they may have had originally. At one place a fragment of green glass from a Japanese fishnet float was found included. It appears that the process of lithification is going on today, but exposed parts of the belt are being eroded—by solution at levels immediately above low tide and by wave action in the higher part of the belt (pl. 5, *B*).

The larger islands of Bikini are located on wide, crescentic sections of the reef and appear to be stable or slowly growing under existing conditions. There is evidence that small islands on the windward side are being eroded on their seaward sides, whereas their lagoon beaches are constantly augmented by foraminiferal sand. These processes result in a slow migration of the islands across the reef toward the lagoon.

LEEWARD REEFS—WESTERN SIDE

Much of the broad leeward reef on the west of Bikini Atoll, specifically the segment that extends from Bokororyuru on the southwest to Namu and near-by islands on the northwest (pl. 1), was not closely examined in many places because none of this 9-mile stretch of reef dries at low tide and there are no islands on which to land. "Landings" on the lagoon side are difficult because of lagoon waves and coral patches; but, on the sea side, deep water extends close to the reef edge. Ships may come in close, and "landings" on the submerged reef can be made with a skiff through the weak leeward surges. The edge of the reef rises a few inches above adjacent parts of the main flat, and the growth of corals and algae that covers it is awash at extreme low tide.

SUBMARINE CLIFFS

The most striking feature of the western reef is the steepness of its seaward slope. This slope was first examined off the westernmost point of the reef by the authors, swimming with face masks; later the profile was measured accurately with a sounding lead from a skiff anchored to the edge of the reef. The visible part of the seaward slope is indeed a steep submarine cliff, as noted by Emery (1948). From a depth of 100 feet to an undetermined depth, greater than 180 feet, the cliff was vertical. The slope appears solid, with a rough, lumpy surface;

PLATE 6

A, Calcareous algae on a lee reef. This columnar type of *Lithothamnion*, growing in colonies about a foot high, is especially abundant on reefs fringing the southern passes and exposed to the waves crossing the lagoon. (Photograph by J. I. Tracey, Jr.)

B, Eroded reef margin, Eninman Island, south side of Bikini Atoll. The marginal zone of this lee reef has been broken by storm waves. The reef flat is narrow (300–500 feet); beyond the island a poorly developed reef extends into Bikini lagoon. Air view looking northwest, taken from about 500 feet altitude. (Photograph by J. I. Tracey, Jr.)

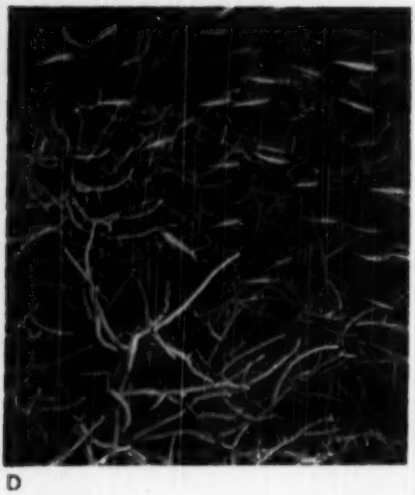
C, Boulder rampart, southeast side of Ourukaen Island. The top of the rampart is about 10 feet above low-tide level and slopes 15°–20°. The seaward edge of the reef, marked by whitecaps, is only 200 feet away. (Photograph by J. I. Tracey, Jr.)

Lithothamnion on a lee reef, eroded lee reef margin and boulder rampart on a seaward reef edge

A

B

C



D

Underwater views on top of coral knoll

some of the rounded masses on its surface may be old coral heads now overgrown with algae. Globular colonies of *Lithothamnion* were seen to depths of 20–25 feet; and green algae, including *Halimeda*, are present. Corals on the steep slopes cover something less than 10 per cent of the surface. There are light-colored patches on small ledges that probably are sand derived from the reef. Leeward cliffs such as those seen on the west side of Bikini Atoll appear to be very stable slopes. They are protected from the prevailing winds and from the heaviest storms by their location.

REEF FLAT

On the western side of the atoll the reef flat is wider than to windward, and it is covered by a greater depth of water. The marginal zone is characterized by a prolific growth of corals and algae, roughly in equal proportions. It rises closer to low-tide level than does the flat behind it; but there is no marginal structure comparable to the *Lithothamnion* ridge of the windward side, and in only a few places are areas of the reef actually exposed at low tide.

The main flat, which exceeds half a mile in width in many places, consists of a hard rock surface, 4–5 feet below low-tide level. The general surface is rougher than on the windward side, as tabular masses of coral 6 feet or more in diameter rise 2–3 feet above the rock flat. Open pools between the coral masses are veneered with coarse coral sand.

The organisms in the marginal zone appear to be well adjusted to existing conditions, and, because of the absence of strong surf, it is doubtful whether the rim can grow appreciably higher. Over the surface of the main flat the growth of corals and other organisms may result in a small net gain, in that the coral areas may be growing upward and slowly extending laterally, thus reducing the pool areas. This process may be aided in some degree by sedimentation, but, as the sand cover is everywhere very thin, this seems unlikely. The fact that on one measured traverse the sand cover on the reef flat thickens slightly toward the marginal rim suggests migration toward the edge of the leeward reef with little or no progressive upbuilding of the reef.

LEEWARD REEFS—SOUTHERN SIDE

MARGINAL ZONE

The reefs of the southern side of Bikini are narrow in most places, and their seaward margins are scalloped or otherwise made irregular by channels and sunken areas. The marginal zone rises about a foot above low tide and is covered by corals and algae. In most places the algae cover the buttresses between channels, the corals and their allies, particularly *Millepora*, occurring in channels and over slopes below low-tide level. The columnar type of *Lithothamnion* is very abundant on the buttresses, and the globular type occurs in the channels; but a *Lithothamnion* ridge comparable to that of the windward reefs is not present (pl. 6, *A*).

PLATE 7

A, Diver standing in thick growth of living coral on top of knoll. (Photograph by K. O. Emery.)

B, Underwater photograph showing school of tuna swimming over coral and sand on top of coral knoll that rises from a depth of 10 fathoms in Enyu channel. (Photograph by K. O. Emery.)

C, Living coral and giant clam, *Tridacna*, on top of coral knoll at a depth of 5 fathoms in Bikini lagoon. Diver's hand indicates scale. (Photograph by K. O. Emery.)

D, Coral thicket at a depth of 10 fathoms in Bikini lagoon. Note school of fish above coral. Base of photograph represents about 4 feet. (Photograph by K. O. Emery.)

Inside the marginal zone a flat ordinarily supports a variety of corals that cover one-half to three-fourths of the surface.

The marginal zone appears to be flourishing and probably is slowly advancing seaward, at least during periods of normal weather.

REEF FLAT

The narrow rock flats of the southern reefs are covered in most places by 2–4 feet of water at low tide (pl. 5, *C*). This depth of water, the narrowness of the reef, and the absence of a marginal ridge that might interfere with circulation permit corals to grow sporadically all over the flat. Most of the colonies are less than a foot in diameter, and the patches on the flat yield the boulders that form the boulder rampart. The flats themselves appear to be in a state of equilibrium, except in those areas where storm waves have eaten deeply into the reef.

ERODED EDGE

On the south side of Bikini Atoll, as shown in plate 6, *B*, there are sharply irregular re-entrants in the reef margin that appear definitely to be due to erosion. The reef edge is cavernous, and, owing to organic growth, the parts near the surface may overhang those below. The edge is composed of living organisms that are adjusted to the small waves and swells that normally strike this coast. Severe storms that at long intervals arrive from the south do great damage to the southern reefs (Emery, Tracey, and Ladd, 1949).

Parts of the edge are thrown upon the reef flat to form the negroheads that are found in that area, and even larger sections may be loosened from the remainder of the reef and may sink to resting places on the shallow terrace that borders parts of the southern reef. The outlines of some of the blocks match those of the re-entrants. The re-entrants thus formed may be 25–500 feet or more wide and may extend into the reef for 25–200 feet. The terrace on which the collapsed sections of the reef edge rest is probably erosional, but it may represent a surface that antedates the present reef.

BOULDER RAMPARTS

Many of the beaches of Bikini Atoll—those along the lagoon as well as those facing the sea—locally are formed of gravel and boulders, but concentrations of these in the form of boulder ramparts are present only on islands located along the south side of the atoll. There are thirteen such islands, nine of which were visited. All those visited had well-developed boulder ramparts along much of their southern coast; on two of the small, westernmost islands the rampart extends along the windward sides of the islands facing passes through the reef at these points; on Enyu Island, which faces the widest pass of all, the rampart is continued around the southwest end and faces the pass on the west; the boulder rampart completely encircles one island, except for a sand beach on the western side.

The ridge of boulders is typically steep (pl. 6, *C*), dipping from 15°–20° seaward in the intertidal zone; above high-tide level the slope is steeper and may be vertical if waves have undercut the rampart. In some places the ramparts extend 5 or 6 feet above high water, descending gently to landward from a narrow crest or rising still higher to landward beyond the break in slope at the top of the rampart.

The boulders vary considerably in size, the largest measuring 2 or 3 feet in maximum diameter. In general, the boulders of the foreslope are somewhat smaller than those of the backslope. Some

boulders are subangular to rounded blocks of reef rock, but most of them are coral heads (*Porites*, *Favia*, etc.) that have been only slightly modified in shape by wave action. Many of the boulders were originally coral colonies much larger than any now growing on the surface reefs, and some consist of forms (such as *Hydnophora*) not now found living on surface reefs at Bikini. It seems probable that many boulders originate in the innominate zone of rich coral growth down the seaward slope, a zone inaccessible to direct observation or dredging.

As previously mentioned, the southern reefs are part of the leeward side but are subject periodically to strong southern swells and on rare occasions to violent storms. Umbgrove (1947) has long recognized that in the East Indies ramparts are highest on the side where the wind exercises its greatest force most regularly. Fairbridge and Teichert (1948), who restudied the rampart system at Low Isles on the Great Barrier, also recognized a relation between rampart development and periodic wind systems. There is not such a rich growth of calcareous algae as on the *Lithothamnion* ridge of the windward coasts, but corals are proportionately more important. The reef flats off the southern islands are narrow—from 200 to 500 feet wide—and their margins are smooth except where broken in places by large erosional reentrants. Under normal conditions at times of high tide waves cross the narrow flats with great force, audibly rolling loose coral boulders over the flat and the seaward slope of the rampart. The ramparts locally form double or even triple ridges that are approximately parallel, each possibly the result of a single storm. In normal times waves that cross the reef tend to flatten the rampart slope, but the ramparts are probably built, moved, or destroyed only by major storm waves.

LAGOON REEFS

The lagoon margin of the main reef, including small reefs fringing the lagoon shores of islands, are described here as "lagoon reefs." The lagoons of the larger atolls, such as Bikini, measure up to 20 miles in diameter and over large areas have depths approaching 200 feet. The reefs and islands of the windward side offer some protection to the lagoon waters immediately in their lee, but, beyond a narrow zone near shore, the winds generate sizable waves, resulting in a continuous circulation of the lagoon waters. This circulation and the constant refreshment of the lagoon by ocean waters through passes and over the reefs are discussed by Von Arx (1948). They are responsible for the rich growth of organisms in the lagoon. The broad swells of the open sea are lacking except near wide passes, such as the one located on the southeast side of Bikini. The lagoon waves engendered by the prevailing trades are primarily responsible for some significant differences that may be noted between lagoon reefs on opposite sides of the atoll.

WINDWARD SIDE

In the lee of Bikini Island on the windward side of the atoll there is a discontinuous series of irregular reef patches that extend from the near edges of the beaches into the lagoon. The only waves that strike these reef patches are the long, slow ocean swells that enter the lagoon through the wide channel west of Enyu Island. These swells, which are probably from the Southern Hemisphere, vary considerably in strength from day to day. They range from swells that rise and fall with little turbulence to

combers more than 6 feet high that break over the reef edge. The reef patches tend to be wider (into the lagoon) than long (parallel to shore), and they show concentric zoning rather than zoning parallel to the shore. Three fairly distinct zones may be recognized:

1. An outer zone with an irregular, lobate margin formed of discrete heads of coral and alcyonarians 4–20 feet in diameter. In the outer part below low-tide level there are large areas covered by hydrocorals (yellow-green *Millepora*, the "stinging coral"), with a few small colonies of true corals. Inside this living margin there are large subcircular areas that uncover a few inches at lowest tide. The rounded outlines of many of these masses have been determined by the alcyonarian, *Heliopora*, and the scleractinian, *Porites andrewsi* (ramose-growth form), but most of the colonies are now dead and veneered with a film of brownish algae. The bottom of the lagoon is covered by sand and broken coral on which there are widely scattered clumps of living *Halimeda*.

2. A middle zone, the floor of which lies 4 feet below low water. Colonies of *Heliopora* up to 6 feet in diameter rise 2 feet above the sand-covered floor, but, like those in the adjoining outer zone, many are dead, and here they are partly encrusted with small, branching corals. The sands of the deeper areas are bound by living *Halimeda*.

3. An inner zone covered by 2–3 feet of water at low tide supports a considerable variety of living corals, but they cover only a small percentage of the bottom, the remainder being sand. The inner zone abuts against eroded beach rock where such is present, elsewhere against the sand of the beach.

Lagoon reefs on the windward side in areas between protecting islands differ somewhat from the type of reef described above. The absence of the island allows the waters of the open sea to wash entirely across the reef at times of high tide —indeed, there is some circulation even at low tide. Where ecological conditions are especially favorable, there may be a fairly rich growth of corals all the way across the reef, from the sea to the lagoon. In other places there may be extensive sand flats or deposits of gravel along the lagoon side of the reef. Over large areas the sand flats are heavily populated with holothurians, whose sand-ingesting activities are responsible for reducing the grain size of the sediment. Beyond, in the shallow waters of the lagoon, there is a growth of corals—chiefly the hardy *Porites* (massive and ramose) and *Heliopora*, with some encrusting algae. The lagoon edge of this zone is deeply scalloped or serrate, with large, irregularly shaped outlying patches. The lagoon reefs of the windward side are probably encroaching slowly on the lagoon—through organic growth and the deposition of sand and gravel.

LEEWARD SIDE

On the leeward side of the atoll the lagoon reefs are better developed than elsewhere because they receive fairly strong waves engendered by the prevailing trade winds in their passage across the lagoon—a distance of a dozen miles. In favorably situated areas—e.g., along the northwest margin of the lagoon—the reefs take on some of the characteristics of the reefs that fringe the atoll to windward. In many places there is a distinct algal ridge bordering the lagoonal reef, though it nowhere becomes the prominent cuesta-like structure that is found off islands that face the sea to windward. The lagoon reefs to leeward are comparatively narrow and do not exhibit the dis-

tinct zonation that characterizes reefs facing the open sea. The coral associations of these reefs lack many of the species characteristic of the windward seaward reefs but include some species found on the leeward seaward reefs. The margins of such reefs in general are irregular, with isolated offshore patches that, like the main reef, are awash at low tide. On the southwest side near passes algae of the columnar type tend to replace the globular colonies, and the hydrozoan, *Millepora*, becomes more important than the true corals, especially on the lagoonal side of the reef. Likewise, near passes, an algal-coral flat may be formed inside the marginal zone, this merging into a rock flat which in many places is strewn with coral-head boulders; some lagoonal shores of the southern leeward islands are bordered with a boulder rampart.

These leeward lagoon reefs do not compare favorably with the thriving reefs that face the sea to windward; indeed, the assemblage of growing organisms is not so impressive as that of the reefs that face the open sea on the lee side. There is, however, a richer growth than on the protected lagoon reefs to windward. The boulder-strewn flats and the ramparts indicate erosion; but over long periods of time the reefs will, under present conditions, encroach on the lagoon and tend to fill it up.

Under particularly favorable ecological conditions the growth of corals on lagoon reefs may be more luxuriant than that on any seaward surface reef at Bikini. The lagoon reef at Latoback Island, Rongerik Atoll, situated on the north side of the atoll is an example of this sort. Coral heads of *Porites andrewsi* and *Platygyra rustica* are of enormous size—larger than any seen at Bikini—as are most of 32 other species observed there.

CORAL KNOLLS

The saucer-like floors of many atoll lagoons, including Bikini and near-by atolls in the northern Marshalls, are made very irregular by the growth of isolated mounds of coral on the bottom. These coral knolls may be quite small—less than 10–50 feet across and 3–20 feet high—but very large ones are numerous, and these may be from several hundred feet to more than a mile in width at the base, rising nearly to the surface of the water from even the deepest part of the lagoon. The 180,000 soundings made in the 24-mile lagoon of Eniwetok Atoll revealed more than 2,000 coral knolls. The Bikini lagoon has not been mapped in as much detail as the lagoon of Eniwetok, but such work as has been done indicates that knolls are equally abundant there. Four of the Bikini knolls were surveyed with considerable accuracy by lead-line soundings and the surface shown by 1-fathom contours. Each of the four was of somewhat different shape, the contoured sides having slopes as steep as 45°; other coral knolls which were investigated with the use of diving gear were found to have vertical slopes for as much as 10 fathoms of their depth.

One of the knolls examined by diving lies about 1¼ miles north of Enirik Pass, on the south side of the atoll. The lagoon floor in this area lies at about 20 fathoms. The knoll rises to within 6 fathoms of the surface. The top is approximately 150 feet in diameter, with a maximum relief of about 15 feet. The surface is rough and irregular, some of the ledges exceeding 5 feet in height. Corals are dominant and cover 75 per cent or more of the area, but their bases are bound by *Lithothamnion* in places, and much of the surface beneath the spreading corals is covered with green, living *Halimeda* and calcareous sand. A great variety of corals is

present, including massive heads of *Porites* several feet in diameter, great fragile brackets of *Acropora reticulata* higher than one's head and 5–15 feet across, and large thickets of staghorn *Acropora* (*A. palifera*, but a form different from that on the surface reefs) in hollows. *Lithothamnion* forms a hard crust or cap over large areas, and white segments of *Halimeda* form much of the sandy debris scattered over the knoll. Mollusks are present, *Tridacna* being the commonest. A number of Tridacnas measuring 1–2 feet in length were found, and one unusually large specimen measured 3 feet. Some of the large Tridacnas were surrounded by growing coral, but others were free and could be easily rolled over. Schools of large fish cruise over the knoll; schools of smaller species hover over single coral heads.

The sand occurs as a filling, or matrix, between corals, and some patches of it are more than 6 inches thick. It consists of unsorted, organic fragments. About half the material is coarser than 2 mm., and of this fraction 80 per cent or more consists of *Halimeda* segments. Coral fragments, mollusk shells and fragments, especially small shells 2–5 mm. long, and several species of discoid Foraminifera form a small proportion of the material.

The sides of the knoll and the edge of its flat top were not examined by diving. The slopes, as seen through a face mask by the authors while swimming on the surface 50 feet above them, appeared to dip off steeply on all sides; it was impossible to estimate the angle of slope accurately, but it appeared to be between 45° and 60°. The side slopes appeared to be white—much lighter than the dark coral-covered top—and were probably covered by sand and coarser debris.

The luxuriant growth of corals and other organisms (pl. 7, *A*, *B*, *C*) on the tops of the coral knolls indicates rather clearly that these structures are growing upward under existing conditions, the tops being raised by corals and algae in place, the intervening areas filled in by organic debris. The sides of the knolls are probably being slowly extended as sand and coarser debris are added to their slopes. The age, origin, and internal constitution of the knolls is not known, as no structure of this kind has ever been drilled. They may represent sporadic growth that was initiated during a part of the Pleistocene, when sea level was much lower than now.

CORAL THICKETS AND *Halimeda* AREAS

The lagoon floor is surfaced primarily by medium and fine sand, *Halimeda* debris, and foraminiferal sand. Coral growth covers only a small part of the lagoon area and is concentrated on the tops of the coral knolls. In some places at lower levels several species of branching *Acropora* form bramble-like thickets such as that shown in plate 7, *D*. These thickets occur in the intermediate depths of the lagoon, where other large areas are covered by *Halimeda* debris or by sand. The deepest parts of the lagoon are covered mainly by *Halimeda* debris and by extensive patches of living *Halimeda*. Near the reefs the sand of the lagoon floor is composed dominantly of "beach" Foraminifera, but the numbers of these decrease rapidly lagoonward, and the sand areas removed from the reefs—especially a wide and irregular belt on the leeward side of the lagoon—are composed primarily of medium and coarse sand.

Very little sediment escapes from the lagoon, and its floor is slowly being raised (1) by debris from the largest areas of

living *Halimeda*, (2) by sand from the marginal reefs and the coral knolls, and (3) by the skeletons of the corals that form the *Acropora* thickets.

SUBMARINE SLOPES

The submarine slopes of Bikini, like those of most of the larger atolls, are very steep near the surface. The average slope on the windward side from the seaward edge of the terrace to 200 fathoms is 35°. On the leeward side, where, as previously noted, the surface reef is bordered by a steep, locally vertical cliff extending to 35 fathoms or more, the average slope from the surface to 200 fathoms is 40°. Below 200 fathoms the slopes flatten progressively until they become tangential with the floor of the ocean at a depth of about 2,500 fathoms. Opposite some of the reef projections (pl. 1) there are broad, rounded spurs that clearly extend to depths below 2,000 fathoms. On its western side Bikini Atoll is connected by a neck with a seamount whose flattened summit at a depth of 700 fathoms is as large as Bikini lagoon.

Some sampling and dredging of the outer slopes was done by Emery in 1946. In 1947 R. Dana Russell, with improved equipment, made a total of 24 hauls from depths of 8–600 fathoms. On the steeper slopes (0–500 feet) the surface is of calcareous sand and gravel partly covered with large blocks of limestone broken off by waves from the edge of the reef. Russell found that the size of the blocks appeared to decrease with depth. Some near the surface appeared to weigh several tons, as a dredge attached by a 1-inch chain to a ½-inch steel cable was lost by breaking of the chain after it had hooked on a large block. The sand between the blocks consists of the shells of Foraminifera, segments of *Halimeda*, fragments of mollusk shells, corals, etc. In the upper 500 feet living corals, algae, sponges, Bryozoa, and other types of marine life were present in abundance.

Russell reported that large blocks seemed to be rare below 200 fathoms, the bottom consisting of limestone boulders and sand, but he expressed the opinion that blocks weighing several hundred pounds extended to more than 350 fathoms because a $\frac{5}{32}$-inch stainless-steel airplane cable carrying a small dredge was broken when the dredge became caught at that depth. The amount of debris encountered at higher levels suggests that the obstruction was a block, but the possibility of bedrock being exposed at that depth must be considered. Below 500 feet there appears to be a considerable decrease in the profusion and variety of marine life. Larger Foraminifera and algae are the most abundant forms, and in this respect some of the dredge hauls from 700 to 800 feet greatly resembled the Tertiary "algal-foram limestone" that is common in parts of Melanesia. With the common forms are a few ahermatypic (nonreef-building) corals, sponges, and brachiopods.

Systematic dredging has been carried out on seaward slopes of Indo-Pacific coral reefs in only a few areas, such as the Maldives (Gardiner, 1898, 1903, Funafuti (David, Halligan, and Finckh, 1904), and Hawaii (Vaughan, 1909), and the data are still far from reasonable completeness for any of these. Therefore, the results of the dredging at Bikini are of considerable interest in their bearing on the problem of the depth to which living reef (hermatypic) corals can grow and on that of changes in coral associations with depth. It has been generally concluded that most reef-building takes place in depths of 15 fathoms or less,

that some hermatypic species extend to 25 fathoms, that a few reach as much as 40 fathoms, and that the maximum depth is finally reached at about 50 fathoms (Vaughan, 1919; and Vaughan and Wells, 1943). Analysis of the vertical distribution of hermatypic types at Bikini adds confirmatory data to these conclusions but suggests that in some localities the lower limit for living reef-type corals may extend to as much as 85 fathoms, although at these depths such corals are scarcely significant contributors to the atoll. From 112 species on the surface reefs at Bikini, the number falls abruptly to about 20 at 15 fathoms, half of these being nonsurface reef species, to about 15 at 45 fathoms, practically all nonsurface species, to about 12 at 70 fathoms, finally dropping to 1 or 2 at 85 fathoms. No hermatypic types were found below this depth, down to more than 120 fathoms.

Below 30–40 fathoms the reef-coral fauna is marked by special species or a facies different from those of the surface-reef forms (*Leptoseris* zone, fig. 2). This is the bathymetric faunal zone recognized first by Gardiner (1903) in the Maldives and later at the same depths in the Hawaiian Islands by Vaughan (1907). At Bikini this zone appears to begin at a somewhat greater depth. It also appears in Bikini lagoon but is not so clearly distinguishable. Ahermatypic corals were found at all depths in small number and variety. Truly "deep-sea" coral types come in at about 80 fathoms and continue on down as far as dredgings extended (in 1947 to 600 fathoms) as the only scleractinians.

The outer submarine slopes of the atoll are being built up and extended at a very slow rate by organic growth at all levels and by sediments derived from the upper levels of the reef.

SUMMARY

It is difficult to evaluate accurately the balance of forces that operate on an atoll such as Bikini because, as has been shown, the assemblages of organisms and the forces of erosion vary greatly from place to place. One fact, however, must be borne clearly in mind: So long as lime-secreting organisms continue to live on the atoll, they slowly but surely are gaining against the sea. As one individual or one group dies and is replaced, the skeletons of the dead forms are added to the growing pile of organically formed limestone that rises from the deep sea. The sea can greatly modify a given section of the reef during periods of heavy weather, the forces of erosion can hold a given section of the reef in equilibrium so that organic growth no more than balances wave destruction, but even in these cases some other part of the atoll gains against the sea: debris is added to the outside talus slopes or is washed across the reef to build the islands or settle in the lagoon. Very little sediment of clay size—sediment fine enough to be carried far from the atoll—is produced.

The process of growth is very slow. Under existing conditions the rigid windward reefs not only prevent truncation—they appear to be slowly encroaching on a pre-existing marginal terrace. Some of the zones behind them appear to be essentially in equilibrium; others are being eroded; but both are contributing sediments that eventually reach the islands or the lagoon as a final resting place. The lagoonal reefs themselves are not to be described as flourishing, but there are large areas of healthy growth which inevitably yield their share of sediment. Only on the south side is the reef rim being slowly eaten away. This reef appears adjusted to normal weather at the present time but not to the severe storms

that strike it at long intervals. Given sufficient time, the storm waves might even cut away the upper reef of this section of the ring. More than half the southern sector is now open to the sea. The Enyu channel is 8 miles wide, the Rukoji pass 2 miles wide; but the shoals that now mark these sections probably never rose higher than they do now. These passes are perhaps best interpreted as parts of an older surface that has been recognized as a terrace elsewhere on the atoll—both outside the existing reefs and in the lagoons.

Bikini Atoll on the whole appears to be a healthy, flourishing structure. There is little reason to believe that growth in the past was markedly richer than now, although corals and their relations may have been proportionately more important in the recent geologic past, when the sea stood 6 feet higher than now.

Continued organic growth—be it slow or rapid—is the one essential requirement for the building of an atoll, but at any given time it appears to be quantitatively unimportant. The bulk of the atoll mass consists of calcareous sediments of varied organic origin, a general conclusion that is supported by all drilling so far accomplished into the foundations of atolls.

REFERENCES CITED

David, T. W. Edgeworth; Halligan, G. H.; and Finckh, A. E. (1904) Report on dredging at Funafuti, in The atoll of Funafuti, Rept. of Coral Reef Committee, Royal Soc. London, London, Harrison & Sons, pp. 151–159.

——— and Sweet, G. (1904) The geology of Funafuti, in The atoll of Funafuti, Rept. of Coral Reef Committee, Royal Soc. London, London, Harrison & Sons, pp. 61–111.

Emery, K. O. (1948) Submarine geology of Bikini Atoll: Geol. Soc. America Bull. 59, pp. 855–860.

——— Tracey, J. I., Jr.; and Ladd, H. S. (1949) Submarine geology and topography in the northern Marshalls: Am. Geophys. Union Trans., vol. 30, pp. 55–58.

Fairbridge, R. W., and Teichert, Curt (1948) The Low Isles of the Great Barrier: a new analysis: Geog. Jour., vol. 111, pp. 67–88.

Gardiner, J. S. (1898) The coral reefs of Funafuti, Rotuma and Fiji: Cambridge Philos. Soc. Proc., vol. 9, pp. 478–479.

——— (1903) Fauna and geography of the Maldive and Loccadive archipelagoes, vol. 1, Cambridge, Cambridge University Press.

Hanzawa, Shoshiro (1942) Coral reefs: Kaiyo no Kagaku (Science of the ocean), vol. 2, no. 7, pp. 470–485.

Kuenen, Ph. H. (1933) Geology of coral reefs: Snellius Expedition, geological results, vol. 5, pt. 2, pp. 1–126.

Munk, Walter H., and Sargent, Marston C. (1948) Adjustment of Bikini Atoll to ocean waves: Am. Geophys Union Trans., vol. 29, pp. 855–860.

Tracey, J. I., Jr.; Ladd, H. S.; and Hoffmeister, J. E. (1948) Reefs of Bikini, Marshall Islands: Geol. Soc. America Bull. 59, pp. 861–878.

Umbgrove, J. H. F. (1947) Coral reefs of the East Indies: Geol. Soc. America Bull. 58, pp. 729–778.

Vaughan, T. W. (1907) Recent *Madreoporia* of the Hawaiian Islands and Laysan: U.S. Nat. Mus. Bull. 59, pp. 32–34.

——— (1919) Corals and the formation of coral reefs: Smithsonian Inst. Ann. Rept. for 1917, pp. 197–199.

——— and Wells, J. W. (1943) Revision of the Scleractinia: Geol. Soc. America Special Paper 44, p. 52.

von Arx, W. S. (1948) The circulation systems of Bikini and Rongelap lagoons: Am. Geophys. Union Trans., vol. 29, pp. 861–870.

SUMMARY OF RUSSIAN PAPERS ON UPPER PALEOZOIC REEFS[1]

MELVIN N. LEVET
California Research Corporation, La Habra, California

ABSTRACT

Several recent short papers on Upper Paleozoic carbonate reefs in the Ural Mountains and North Russian Plains are summarized. Reef migration and relation to tectonic activity, petrographic features and dolomitization of the Ishimbaevo oil-bearing massifs, and lithology of small reefs exposed on the North Russian Plain are described briefly.

REEF MIGRATION

The migration of Upper Paleozoic reefs of South Bashkiria is described by Theodorovich (1941). These reefs are confined to the Upper Carboniferous and Lower Permian. The reef belts are in general parallel to the main strike directions of Carboniferous deposits of the western slope of the South Urals.

The oldest reefs (Zilim) of the Upper Paleozoic are situated in the most easterly zone, namely, in the frontal ranges of the western slope of the South Urals. These reefs are less than 40 meters thick and in typical cases are very pure, being characterized by the nearly complete absence of dolomite. The next younger reefs (Gzhel stage) are situated in a more westerly belt, namely, in the piedmont part of the western slope of the South Urals. These reefs, now up to 50 meters thick, were apparently subjected to erosion during Lower Permian time. Commonly there is a [direct] relation between dolomitization and the quantity of disseminated FeS_2, which has been mostly transformed to iron hydroxides. Apparently, dolomitization proceeded in the presence of decomposing organic matter. The youngest reefs of the Upper Paleozoic (Shikhan reefs of Sakmara and lower Artinskian stage) are situated in a still more westerly belt. These reefs have a greater thickness than the preceding reefs. They also have undergone erosion. Still farther westward are the buried reef massifs of the Ishimbaevo oil-bearing region, which are comparable in age with the Shikhan reefs. They are very thick, in spite of erosion during the later portion of the Artinskian age. Dolomitization is common.

A review of the distribution of the reefs of the Upper Paleozoic of South Bashkiria reveals a close relationship between their chronological appearances and duration of existence and the tectonic history of the South Urals. As the Hercynian folding of the South Urals proceeded, the reef structures shifted from east to west in accordance with the migration of the Cis-Ural down-warping in the same direction.

CHARACTERISTICS OF THE ISHIMBAEVO MASSIFS

Petrographic studies of cores by Theodorovich (1942*a*) of the Ishimbaevo massifs demonstrate their reef origin.

The Ishimbaevo massifs consist mainly of limestones made up of whole skeletal remains of such organisms as bryozoans, hydractinoids, brachiopods, calcareous algae, enveloping calcareous sponges, and fusulinids. The great quantity of biomorphic and biomorphic-detrital limestones and the type of organisms

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of which they are composed are evidence of a reef origin. The limestones are massive, unstratified structures which are very porous and cavernous as the result of incrustations. The irregular relief of the limestones composing the Ishimbaevo massifs, as well as the presence of fusulinids at horizons higher stratigraphically than normal, suggests a reef origin. Absence of terrigenous material is consistent with a reef origin.

In addition to the typical reef limestones, the Ishimbaevo massifs contain some detrital material which may have been deposited both on the periphery of the reef and in small, more or less central hollows of its elevated part. Of all the calcareous formations of the massifs, the detrital facies are the most favorable for dolomitization and anhydritization. The calcareous oil reservoirs of the Ishimbaevo massifs are mostly detrital structures. The origin of the massifs is believed to be associated with the appearance of meridional gentle folds formed below sea level between Upper Carboniferous and Lower Permian time. Folding contemporaneous with reef growth affected, to some extent, the structure of the reef massifs. The growth of the massifs was apparently accompanied by a slow subsidence of the corresponding part of the sea bottom.

The algal reefs of the Sterlitamak region in the Urals (Sulin and Varov, 1933) contain two fundamental types of carbonate rocks. The first type is characteristically "rubbly and silty" and was formed by the accumulation of eroded organogenetic sediments in the voids of the reef. The second consists of organogenetic constituents cemented in place by calcite. Fossil assemblages of these types are aphanitic and encrusted limestones, respectively. The aphanitic types contain 0.92–4.22 per cent insoluble residues, whereas the encrusted limestones contain only 0.14–0.50 per cent insoluble residues. Lithologic and petrographic examinations show algae to be the basic rock-forming constituent. Crinoids, sponge spicules, and brachiopods occur in lesser numbers. An observed increase in insoluble residues with a corresponding increase in dolomite content suggests a common origin of these features.

ASPECTS OF EXPOSED REEFS

Zekkel (1941) has described reefs of the Kazanian stage in the North Russian Plain. In certain localities strata associated with the reefs are not horizontal but appear to form anticlinal folds, the beds dipping as much as 16°. The reefs themselves attain a height of only 7 meters, but, inasmuch as the roots are not exposed, this is probably only the topmost part. An associated stratum of considerable extent has an abundant bryozoan and brachiopod fauna. Nearer the region of reef occurrence this stratum is thicker and the dimensions of the enclosed bryozoan skeletons increase considerably. It is possible that this stratum marks the time when reef growth began. If this is so, the actual height of the reefs is about 35–40 meters.

In the neighborhood of buried reefs there commonly occur thick seams of diagonally stratified (cross-bedded?) brownish limestone consisting almost solely of separate members of crinoids and, in addition, worn shells of thick-walled *Athyris* and unrecognizable bits of other organisms. White limestones very rich in large forms of corals, bryozoans, brachiopods, pelecypods, and especially crinoids are commonly found in the immediate vicinity of the body of the reef. Evidently conditions were favorable for the growth of these forms, although the greater part of the fauna

was destroyed and comminuted by abrasion. Moreover, strong and ever changing currents in the region of the reefs carried away fossil remains, as evidenced by the seams of crinoidal limestone. On the other hand, those forms not destroyed or washed away, which probably lived in relatively closed lagoons, were buried in their dwelling places and make up the white limestone. The reefs made possible an apparent intermixing of the lower and upper Kazanian faunas. Thus, the presence at relatively high levels of *Spirifer* fauna, characteristic of the lower Kazanian, is explained by the fact that the strata bearing that fauna normally occupied a lower level but, where deposited on the sides of the reefs, occupy higher levels.

DOLOMITIZATION IN THE ISHIMBAEVO OIL REGION

Theodorovich (1942*b*, *c*) has described the process of dolomitization in the reefs of the Ishimbaevo oil region. Cores of Upper Paleozoic reef rocks show that dolomite is commonly associated with anhydrite and in some places with gypsum. Petrographic data indicate shallow waters and show that both dolomitization and anhydritization were extensive. Also, there is evidence of the existence of very shallow bays or lagoons and insular areas. Emergence alternated continuously with submergence. Dolomitization began on a very small scale and was not accompanied by precipitation of calcium sulfate. It took place either previous to or simultaneous with incrustation and partial recrystallization, producing small areas of extremely fine-grained dolomite which replaced fine-grained calcite around the skeletal remains of Foraminifera, calcareous algae, etc. Most dolomitization occurred later and was accompanied by anhydritization. Oozy, detrital calcareous deposits were subjected to selective replacement at the bottoms of small bays or other water reservoirs at the reef's surface. At the time when portions of the reefs were emergent as insular masses, primary dolomite may have been deposited as incrustations in fissures or on corroded surfaces of shallow depressions. Finally, when the reefs were submerged by the concentrated water of the basin, all the freely accessible cavities of the reef complex were rapidly filled with fine-grained anhydrite. At the boundary between the limestone and fine-grained anhydrite, thin, discontinuous zones and aggregates of very small grains of primary dolomite were formed. In places small amounts of crystalline silica occur along the periphery of the anhydrite. Perhaps the higher parts of the elevated reef were islands, and the silica may have been the product of long-continued continental weathering of these insular portions of the reef. It was probably carried into central depressions of the elevated portion of the reef as colloidal solutions. (Presumably the source of silica was siliceous sponges and/or terrigenous clastics incorporated in the reef.)

REFERENCES CITED[1]

SULIN, V. A., and VAROV, A. A. (1933) Oil-field limestone reef facies in the Ural: Neftyanoe Khozyaistuo, Petroleum Industry, Moscow, vol. 24, no. 1, pp. 21–24. (In Russian.)

THEODOROVICH, G. I. (1941) Migration of Upper Paleozoic reefs in South Bashkiria: Comptes rendus (Doklady) de l'Académie des Sciences de l'U.R.S.S., Moscow, new ser., vol. 32, no. 7, pp. 499–502.

[1] All references are on file in the Library of Congress, Washington, D.C. With the exception of the paper by Sulin and Varov, all references are in English.

——— (1942*a*) On the nature of Ishimbaevo massifs and the prospecting of new oil fields of this type: *ibid.*, vol. 34, no. 3, pp. 93–96.

——— (1942*b*) Dolomitization of reefogenic formations in the oil-bearing region of Ishimbaevo: *ibid.*, no. 6, pp. 160–164.

——— (1942*c*) Process of dolomitization in the eastern massif of the Ishimbaevo oil-bearing region: *ibid.*, no. 7, pp. 199–201.

ZEKKEL, J. D. (1941) Importance of reefs in the stratigraphy of the Kazanian stage: Comptes rendus (Doklady) de l'Académie des Sciences de l'U.R.S.S., Moscow, new ser., vol. 32, no. 8, pp. 569–571.

NIAGARAN REEFS OF THE GREAT LAKES AREA[1]

HEINZ A. LOWENSTAM
University of Chicago

ABSTRACT

The controversial issue as to whether or not certain structures in the Niagaran of the Great Lakes region are true reefs is reviewed. Reefs are defined in terms of fundamental biologic potentials of the organisms responsible for them rather than in terms of the present-day appearance of the structures. These potentials are the ability to erect rigid topographic structures by frame-building, sediment-retention and binding, and to create a wave-resistant structure. These have been the basic properties of reef-building biota throughout all time.

Organisms potentially capable of erecting the Niagaran structures are shown to have been present. Evidence of their performance capacity is presented. The biologic affinities of the common builders, combining frame-building and sediment-binding abilities, remain to be determined. Stromatoporoids performed these functions in some cases. The character of the niche-dwelling assemblage is reviewed in terms of environmental relations and potential source material for bioclastic debris. The fossil assemblage, the growth forms, and the character of the sediments both off and on the reef enable one to distinguish quiet-, semirough-, and rough-water environments. These criteria show that the builders raised the structures from quiet-water bottoms into the agitated surface waters. The structures are shown to be wave-resistant. Ecologic zoning cannot be demonstrated because of the removal of the critical sections of the core that embodied corresponding growth surfaces and because of masking action by dolomitization. Further evidence of the reef nature of the Niagaran structures is provided by bioclastic debris fans composed of richer and more varied assemblages as compared with those of previous growth stages, structure-mantling debris covers, the halo of reef-created turbulence, and the contribution of much fine reef outwash to the archipelago bottoms. The Niagaran structures, therefore, exhibit all the basic properties of true reefs.

The reef shapes, the direction of maximum reef expansion, and the location of the zone of settling of fine clastic sediment by-passed around the reefs show the prevailing winds to be southerly. An example of mass destruction on the lee of a wave-resistant reef, resulting from a northwesterly storm, supports this view.

The substrate from which the reef rose is shown to have been unconsolidated muds and bioclastic sands. Coarse skeletal debris is assumed to have formed the attachment surfaces for the pioneer population, but evidence is lacking because of the inadequacy of exposures.

The character of the surrounding deposits indicates the probable depth of water; thus the heights attained by the reefs can be estimated. This method shows that the reef heights, ranging from less than 10 feet to a maximum of about 200 feet, increased progressively from the clastic-free belt southward to the archipelago border.

The pattern of Niagaran reef distribution is only faintly discernible. Reef-building on a small scale began in Albian time in Michigan and Ontario. Most vigorous growth and maximum spread, resulting in an impressive archipelago, occurred in late Niagaran time. The reef complex progressively expanded west of the Cincinnati arch from a more central position toward its ultimate southern border.

High calcium metabolism, shown by mass fixation of the reef biota, and the development of evaporite basins, which may have started inside the archipelago during Niagaran time, constitute evidence of a moderately warm climate in the Niagaran reef archipelago. Tilting of the earth's axis or continental drift, thought by some to account for the spread of reefs into northern latitudes, is not necessary; vigorous northward movement of warm waters due to prevailing southerly winds seems all that is required. A notably lower temperature threshold for the Niagaran reef-builders is also possible.

INTRODUCTION

The Niagaran reefs of the Great Lakes region have been much studied ever since Hall (1862) interpreted the fossil-rich structures of Wisconsin as reefs. Progress has been made in accumulating data leading to the understanding of their origin and environmental relations, despite inadequate exposures and large-scale dolomitization (Chamberlin, 1877; Grabau, 1903, 1924; Cumings and Shrock, 1928; Fenton, 1931; Bretz, 1939; Shrock, 1939; Lowenstam, 1948*a*). The discovery of subsurface reefs in the last ten years in Illinois and Indiana has greatly increased our knowledge of their areal distribution. Although lacking the detail provided by outcrops, the subsurface records provide a three-dimensional

[1] Manuscript received April 20, 1950. This paper is based in part on studies made for the Illinois State Geological Survey.

picture of the reefs that contributes much toward an understanding of their environmental setting (Lowenstam, 1949*a*). In turn, the understanding of regional environmental relations has greatly aided in understanding the observed differences in reef forms and the surrounding sediment character of the widely separated outcrop areas. We can now secure more specific criteria for determining the growth behavior of the reefs and their environmental relations in their respective habitats.

The abundance of corals and stromatoporoids; the suggestion of calcareous algae; the fact that the structures were the sites of abundant life, as witnessed by the abundance and variety of large skeletal remains in contrast to the comparatively meager, fragile fossil remains of distinct assemblages in the enclosing strata; and, finally, the structural expression and the flanking detrital fans led to the general acceptance of these structures as true reefs. That a large amount of detritus was contributed by these structures to adjacent bottoms has been established by recent subsurface studies and by evidence of zones of turbulence around certain structures (Lowenstam, 1948*b*, 1949*a*). Yet Roll (1934) considered the Niagaran reefs of North America and Gotland the genetic equivalents of the sponge banks of the Upper Jurassic of southern Germany and denied that these Niagaran structures were true reefs. Similarly, Lecompte (1938), in the first critical comparison of the Niagaran and Devonian structures of North America with present-day reefs, came to the conclusion, principally on biologic grounds, that the Niagaran structures differed in so many essential aspects from the present-day reefs that he seriously questioned their reef nature. Because Roll based his conviction on published data, his criticism carries less weight than that of Lecompte, who briefly examined the Niagaran structures of Niagara gorge and the Wabash and Bluffton reefs in Indiana. Because Lecompte's criticisms are based primarily on biologic evidence, the very essence of the reef problem, it is not possible to ignore his view that the Niagaran reefs are biologically unlike modern reefs.

It is the purpose of this paper to review critically the fundamental qualifications of the Niagaran structures as reefs in the light of the information derived from a recent subsurface study (Lowenstam, 1949*a*) and subsequent field observations. This study is limited essentially to the reefs in the southern portion of the Great Lakes region, although discussion of certain aspects of the problem is extended to the presently known distribution area of the Niagaran reef archipelago. Speculative in part, it is designed to focus attention on the problems which require further study.

REEF DEFINITION

Fundamentally, reefs are expressions of biologic activity and must be recognized as such not only by neobiologists but also by paleobiologists, stratigraphers, sedimentologists, and geomorphologists. As reefs are the products of highly specialized aquatic communities, distinguished by a combination of potentials inherent in other communities, where they are, however, partially masked by the habitat, they must be approached primarily from an ecologic point of view. Only by defining reefs in terms of ecologic potentials can we properly understand their relation to other genetically related communities. It then becomes possible to make a clear and consistent distinction between reefs and both related and unrelated structures in

the geologic record. Only then can we clearly survey and evaluate the evolutionary changes of reef-building organisms and communities of the past.

At first glance it seems unnecessary to repeat what is known to essentially all who have been engaged in fossil reef studies, as embodied in the many reef definitions found in recent geologic literature. Yet we find in many reef studies that, once we are past the definition, the evidence is too often so inadequate that the reader remains in doubt as to whether or not the author was dealing with true reefs. Ladd's recent critical review of reef studies (1944) amply documents this situation. There is an unfortunate tendency to classify any carbonate lenticle or localized carbonate thickening as a reef or reeflike structure. The problem is aggravated by the chemical instability of pure carbonate aggregates, which are prone to alteration by recrystallization and dolomitization, which may completely mask original textures. Yet, unless we adhere rigidly to an accepted reef definition, we shall create confusion instead of making progress. If we are confronted with completely altered structures which, with means now at our disposal, cannot be shown to be reefs even though they strongly resemble reefs in form, it would be better to state explicitly their problematic nature instead of assigning them to the convenient catchall category of "bioherm." It is rather curious that the term "bioherm," specifically coined by Cumings and Shrock (1928) to achieve refinement in definition, has become a receptacle for a multitude of carbonate lenticles which are wholly unrelated in origin. The reason for this appears twofold. In the first place, any lenticle or vertically extensive bank-shaped body composed principally of massive carbonate and clearly distinguished from its enclosing strata is unhesitantly classified as a bioherm, particularly if it is notably fossiliferous and contains corals, algae, bryozoans, or even crinoids or vagile pelecypods. It is apparent that its present structural appearance is the chief criterion of origin, despite the fact that Stockdale (1939) has shown that the Borden bioherms were, at the time of their formation, only patches of dense crinoid growth which at no time stood appreciably above their surrounding bottoms. Because these communities maintained their position in time, they formed aggregates or accumulations that now appear as reeflike bodies. The same is true of the *Lucina* buttes cited by Cumings (1932). Surely, any paleontologist or geologist who came across patchily distributed crinoid meadows or *Lucina* banks in the sea today would not associate them with, or refer to them as, reefs. Yet, despite understanding of this point, Cumings has listed them as examples of bioherms, primarily because they form lenticular bodies. As soon as the same assemblages, even corals, occur in distinctly bedded deposits, they constitute biostromes, not reefs. Roll (1934), in his study of the Upper Jurassic "sponge reefs" of southern Germany, showed that they originated as a succession of sponge banks and, although their origin and relation to their environment were shown to be radically different from true reefs, they were nevertheless retained by him as a distinct type of reef simply because of their present-day structural appearance as such.

The second reason for the confusion arises from a lack of critical biological analysis and of understanding of the biologic properties of the organisms supposed to have built the structures. The presence and abundance of corals, algae, and bryozoans as such do not constitute

proof of reef-building activity. As we go further back in geologic history, we must examine each case critically to determine whether or not the organisms in question were potentially able to erect structures and to bind sediments and, if so, whether or not they actually did. Ladd's contention that the term "reef" should be applied only to fossil structures in which organisms not only exhibit reef-building activity but also, as he stresses, show evidence of wave-resistant qualities is fully justified in the light of the general lack of documentation in the reef studies reviewed.

Reefs of today are well-circumscribed biological entities. If we formulate the principles of reef-building in terms of potential, we arrive at an ecologic definition of reefs which should apply equally to reef communities of the past and those of the present, even though the reef-builders of different times were totally unrelated. This implies a shift of emphasis from the structure to the structure-building potential. By strictly adhering to the principles not only in theory, as has been generally agreed, but also in practice, the term "reef" can and will be useful, a point in which the writer fully agrees with Arkell (1933).

What, then, are the characteristics and potentials which distinguish the reef community from other biota? The frame-building and sediment-binding organisms have the ability, irrespective of combined or divided functions, to erect their own hard substrate and the potential actively to raise their own foundation, in the form of discrete topographic structures, from the depth of quiet waters into the turbulent surface waters. Thus they create their own shore habitat which they are capable not only of maintaining but also of expanding in the wake of destructive physical forces. Once having attained wave resistance, the reef habitat expands its influence to its surroundings, locally affecting circulation and salinity and effecting sedimentation actively through contribution and passively through creation of a zone of turbulence. Thus, instead of total control by the environment, we find, within certain limits, independence, mastery, and even control of the reef habitat by the organisms which created it. The wave-resistant qualities of the reefs make them major local sources of carbonate materials, and they also affect the sedimentation conditions of their surroundings over a considerable area.

Thus a reef, in terms of ecologic principles, is the product of the actively building and sediment-binding biotic constituents, which, because of their potential wave resistance, have the ability to erect rigid, wave-resistant topographic structures. These are the principles that are common to both past and present-day reefs but which leave room for specific evolutionary adaptation.

Reefs defined in terms of the ecologic potentials of their biota are clearly set apart from unconsolidated banks, although certain sponge, pelecypod, brachiopod, and (in the past) crinoid banks may be topographically well-defined, low-lying structures, the sites of intense organic activity. Nevertheless, these are biotically clearly differentiated from reefs, and, in terms of their ecologic potentials, they fall well outside the limits of true reefs. This has been clearly demonstrated by Roll (1934) for the Upper Jurassic sponge banks of southern Germany, which were totally controlled by their environment.

Bank-forming organisms with compound skeletons, even if sedentary, are incapable of raising their own substrate, as their skeletons disarticulate more or

less rapidly after death. They are only an abundant source of skeletal debris, which, in quantitative terms, greatly exceeds the carbonate contributed by the biota of the surrounding areas that have a lower population density. The differential rise of certain bottoms above their surroundings is thus primarily a function of differential rates of accumulation of bioclastic debris. As such, the organisms play a passive role in bank-building. Lacking debris-binding elements, such banks are bounded by low-angle slopes, characteristic of unconsolidated deposits. Upward growth of the bank surface can take place only below effective wave base, which is the ceiling of upward expansion. Once near wave base, such banks undergo intermittent sweeping and sorting action by storm-induced turbulence. Such action leads to marginal expansion by accumulation, which, in the fossil record, may simulate reef-flank growth, as has been pointed out by Roll. A critical point is that such unconsolidated banks cannot be sources of carbonate deposits, as true reefs are. They lack the biotic mechanism to penetrate into the surface waters, and, above all, they lack the regenerative power to deposit carbonate in excess of the loss caused by destructive physical forces. Thus their influence on other habitats is essentially confined to their own periphery. If such banks, maintained at effective wave base, are located within the area of deposition of fine, clay-sized terrigenous clastics, they contrast sharply in character with the deposits of the surrounding area. Reduction of the rate of settling of the terrigenous clastics, together with dilution by rapid accumulation of skeletal debris, reduces the ratio of terrigenous muds. Intermittent, increased turbulence may disperse the mud and carry it into the surrounding bottoms. Sorted bioclastic debris, heavier than the carrying capacity of the agitated waters, may accumulate on the banks during relatively short storm periods.

Occupation by sediment-binding organisms may take place, as has been shown by Walther (1927, p. 60) to be the case on the Pigeon bank in the Gulf of Naples. Within an interval of twenty-five years, this bank, originally covered by a fine Foraminifera-bearing shell sand, became fully encrusted by a heavy cover of calcareous algae. Similar conversion from soft- to hard-bottom banks or even reefs may, through faunal succession, thus take place intermittently, depending on the degree of stability of the substrate. Careful analyses of reeflike structures in the geologic record must be made to determine whether we are dealing with a nonreef, soft-bank accumulation or a composite structure of soft, mud-bottom banks and reefs.

Hard-bottom banks, e.g, those made by calcareous algae, or the tropical rocky-shore habitats characterized by a combination of sediment-binding algae, cementing Foraminifera, pelecypods, gastropods, and barnacles, exhibit essentially the reef-building potentials, some including the critical wave-resistant qualities, embodied in these biota. Rocky-shore biota of tropical seas have a definite ceiling at a shallow depth, and for this reason they occupy reefs only where these are at, or slightly antecedent to, the wave-resistant stage. They cannot form and raise reefs from bottoms that lie well below effective wave base. If their substrate is actively sinking, it is not quite clear whether they alone could expand, rather than maintain a balance between growth and destruction, to form reefs without reef-building organisms such as the Madreporaria of today. Upward growth as controlled by photo-

tropism and rapid growth rates prior to emergence in the surface waters seem to be two essential attributes of the reef-building corals that make possible rapid upward growth of reefs and their maintenance on actively sinking bottoms. It would thus appear that, without elements of such abilities, banks formed exclusively by cementing forms, such as calcareous algae or oyster banks, the latter not wave-resistant, could not form true reefs. Although closely related to reefs, they appear to constitute, instead, links in the grading spectrum of allied habitats, as is to be expected in nature, rather than true reefs. This as yet fluid boundary points to the need for focusing attention, in fossil reef studies, on the actual performance record of potential reef-builders on a regional scale.

THE QUESTION OF THE REEF NATURE OF THE NIAGARAN STRUCTURES

To qualify as reefs, the Niagaran structures must show evidence that they were erected and strengthened by organisms capable of building and raising a skeletal framework above the sea bottom and potentially capable of growth into turbulent surface waters, to which they would be wave-resistant. The crucial point is the presence or absence of reef-building and sediment-binding organisms. If these are present, questions dealing with whether or not they actually attained wave resistance, with their reef-building potentials, with the nature and role of the reef dwellers as a potential source of bioclastic debris to facilitate rapid reef expansion, and with the manner in which the reef communities coped with specific environmental factors, become germane. From these and by comparison to those of the present day, we may deduce the specific ways in which the Niagaran reef biota gained mastery and control of their environment.

Roll (1934) believed that there were no true Silurian reefs but only unconsolidated bank-type accumulations, whereas Lecompte (1938) believed that the so-called "reefs" were only deeper, quiet-water organic mounds populated by communities not only radically distinct from present-day reef communities but also potentially incapable of reef-building. To decide the issue, the basic aspects will be critically and individually examined in terms of their bearing on individual points of dispute.

REEF-BUILDERS

In a period as remote as the Silurian, we find organisms whose phylogenetic relationship to those of the present day are very distant, uncertain, or nonexistent. Most important is the fact that their reef-building ability and ecology are a priori unknown. It is deductively logical and has been a common practice to consider representatives of the classes and orders that embrace the present-day reef-builders, such as colonial corals and calcareous algae, as performers of similar functions in the past, particularly if they have similar morphology and indicate similar facies relations. Thus Vaughan (1911), in the first general survey of Paleozoic reef ecology, came to the conclusion that

> the same groups of reef building organisms are represented in both the Paleozoic and Recent seas; in both they have the same facies as regards growth-form; in both their physiologic activity has resulted in the secretion of large quantities of carbonate from the surrounding sea water, and in both submarine banks, known as coral reefs have resulted. The general similarity of the organisms and the similarity in the result of their physiologic activity assuredly suggest similarity of conditions under which the physiologic process took place.

This has been the general approach of all who considered the Niagaran structures as true reefs. As to the question of Niagaran reef-builders, the relative abundance of stromatoporoids, tabulates, and colonial tetracorals commonly found in

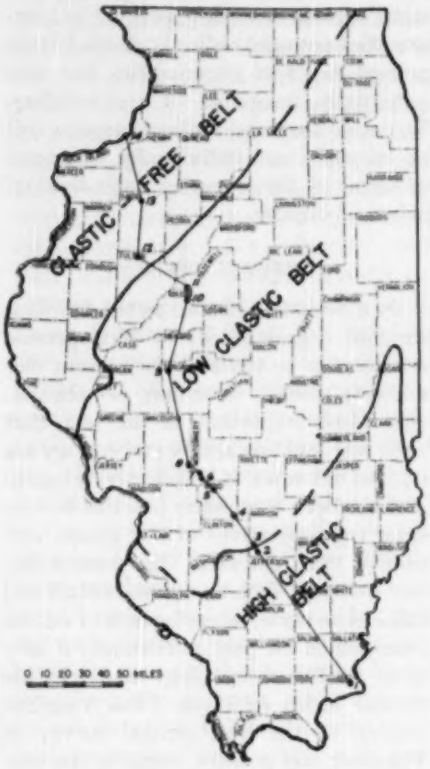


FIG. 1.—Niagaran sedimentation belts in Illinois, showing line of cross section (fig. 2).

those structures in which dolomitization has not entirely obliterated the impressions of organic remains, particularly in the commonly developed, steeply inclined flanking beds, has been generally accepted by the contenders of their reef origin as evidence that these forms were the principal Niagaran reef-builders. They were possibly aided by calcareous algae (Cumings, 1932; Shrock, 1939), although the presence of these forms has not as yet been definitely established. If we examine critically the criterion which establishes these forms as reef-builders, we find it to be, aside from biologic affinities and general similarity in gross morphology to present-day reef-builders, the localization of their greatest numerical and varietal abundance in the Niagaran structures. These facies-linked concentrations are most strikingly exhibited where contrasting facies border and enclose the structure, as is commonly the case in the low-clastic belts (Lowenstam, 1949*a*; and figs. 1, 2). Here the enclosing deposits, averaging 15–20 per cent terrigenous clastics (locally as much as 50 per cent), generally carry few corals. Those found in the clastic-rich deposits, if conspecific with colonial forms in the structures, are generally small, differ in growth morphology, and are commonly numerically subordinate to solitary forms not found on the structure. However, as soon as pure carbonate deposits extend as tongues from the structures into the bordering sediments or form individual units within them, colonial corals and particularly stromatoporoids become more abundant, and differentiation on the basis of this criterion becomes a matter of degree.

The only definite conclusion which may be drawn from these facies-related distributions of the stromatoporoids and corals is that, because the seas in the area of the clastic belt carried terrigenous clastics in suspension, the environmental factor responsible for the by-passing of sediments, that is, agitated waters, favored their development, but it does not follow that they were necessarily reef-builders. They may have merely formed fairly dense, meadow-like carpets, which did not rise at all or rose only slightly above

their surrounding bottoms. Such is the case in the "coral beds" of the Manistique formation of the Northern Peninsula of Michigan, where corals seem fairly densely crowded, with individual colonies reaching several feet in diameter but showing no tendency to fuse laterally. The well-bedded character of the matrix further precludes confusion with reefs. The lenticular to mound shape of the coral-bearing structures has been used as evidence that the stromatoporoids and colonial corals actually were potentially capable of reef-building and did erect these structures above the surrounding bottoms at any given time during their development. But may not these have been only low-lying stromatoporoid-coral banks, of limited areal extent at best, which continued to grow uninterruptedly

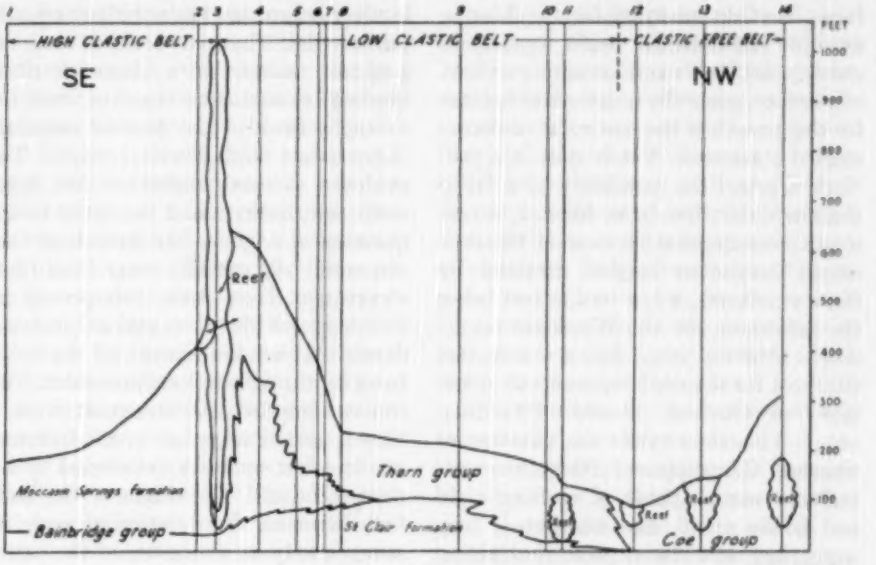


FIG. 2.—Diagrammatic cross section of the Niagaran strata from southeastern to northwestern Illinois along line indicated in fig. 1.

WELLS IN THE SECTION

No.	Company	Farm	Sec.	Twp.	Range	County
1	Marietta	No. 1 Fricher	30	11 S.	8 E.	Hardin
2	Texas	No. B-9 Carr	7	1 N.	2 E.	Marion
3	Martin	No. 1 Robinson	4	2 N.	1 E.	Marion
4	Texas	No. 1 Ensloe	6	4 N.	3 W.	Bond
5	Kingwood	No. 1 Gaffner	30	6 N.	2 W.	Bond
6	Reznick	No. 1 Monke	4	7 N.	5 W.	Montgomery
7	Seaboard	No. 1 Lay	20	8 N.	5 W.	Montgomery
8	Fuller and Turner	No. 1 Chamness	4	9 N.	6 W.	Macoupin
9	William and Whittinger	No. 1 King	24	15 N.	7 W.	Sangamon
10	Hayes	No. 1 Null	19	22 N.	6 W.	Mason
11	Pekin	No. 1 De Sutter	8	23 N.	6 W.	Tazewell
12	Young	No. 1 Midland Electric	2	8 N.	3 E.	Fulton
13	Thorpe	No. 2 Little John Coal Company	36	12 N.	3 E.	Knox
14	Thorpe	No. 5 East Moline State Hospital	19	18 N.	1 E.	Rock Island

and maintained essentially their position in space? It is thus that we are confronted with the bank versus reef question, discussed in general terms in the introduction, specifically raised by Roll (1934) for the Silurian reefs, and implied, although not specifically worded as such, by Lecompte (1938). The facies relations noted above and the evidence of a zone of turbulence around several of the more recently established subsurface structures, best documented for the Marine example (Lowenstam, 1948*b*, 1949*a*), do not support Roll's and Lecompte's views of a deeper, generally quiet-water habitat for the growth of the lenticular or dome-shaped structures. Yet it does not preclude a priori the possibility of a fairly shallow-water, low-bank habitat, a reasonable assumption in view of the estimated maximum heights attained by these structures, a few tens of feet being the maximum for the Wisconsin occurrences (Shrock, 1939) and a few meters the most for the contemporaneous structures of Gotland, Sweden (Hadding, 1941). This then raises the question of whether the supposed Niagaran reef-builders were capable of erecting rigid reef bodies at all, and whether, if they were, they were also capable of attaining and coping with wave resistance. The Swedish structures are shown by Hadding (1941) to be formed by compact frameworks of algae, colonial corals, and stromatoporoids above effective wave base, as corroborated by bedded bioclastic debris outwash. Yet inability to attain wave resistance is clearly implied, as Hadding states that "the reefs of Gotland were, however, not formed immediately below ebb level but at a somewhat greater depth, estimated at 5 to 50 meters. At an elevation of land the reef itself could grow 20 meters high under certain circumstances at the same time as the bottom was equally raised." In North America the mode of skeletal frame-building activity has, up to the present, not been considered. It is curious to note that the only documented record of frame-building activity is that from the Wabash structure, noted by Lecompte, who denied the reef nature of these structures.

The question of wave resistance achieved by cementing and sediment-binding organisms has remained equally unanswered. The sole record of such organisms consists of a single detritus-binding colonial coral reported from the southern flank of the Marine structure (Lowenstam and Dubois, 1946). The evidence of wave resistance has been, until now, indirect and is related to the question of origin of the flank beds that commonly dip radially away from these structures. These were interpreted by Cumings and Shrock (1928) as bioclastic debris that has been swept off the structures by the agitated surface water. This concept implied wave-resistant reefs, a view generally accepted by all American students but seriously questioned by Lecompte (1938). The origin of the flank beds bears on the question of wave resistance only as a measure of the extent to which Niagaran reef-builders were capable of mastering the turbulent surface waters. A prerequisite to this view is evidence of organic adaptations to the task. The uncertainty concerning the origin of the Niagaran structures stems basically from lack of information on the reef-building potentials of the organisms involved—a question which one would expect to have been of prime importance in a case of organic activity.

To prove that the structures could have been raised as distinct reef bodies above their surrounding bottoms requires adequate documentation of skele-

tal frame-building to assure their stability in the agitated surface waters or, if the reefs started to grow upward from deeper, quiet-water bottoms, the ability actively to maintain themselves demands sediment-retaining and, above all, sediment-binding organisms.

Examination of a number of adequately exposed structures has revealed clear-cut evidence of frame-building activity. The most instructive example of extensive framework observed to date occurs in the Wabash structure, which is excellently exposed in the railroad cuts just north of the Big Four station at Wabash, Indiana. Cumings and Shrock (1928), who described the Wabash structure in detail, stated that the massive core consists chiefly of a diagenetically altered breccia or sand of stromatoporoids and other organic fragments. Lecompte (1938) was the first to point out and illustrate the lacy network of crystalline ribbons in the core. These, said by him to be devoid of organic structures, did not resemble diagenetically altered stromatoporoids but showed a surprising resemblance to problematic structures of the Middle Devonian reefs of Belgium, which had been referred to *Stromatactis*. Lecompte also pointed out that the *Stromatactis* network extends past the massive core on the northern flank into what Cumings and Shrock considered the proximal portions of the flank beds.

The writer found that this framework extends irregularly down the flank beds, past the portions indicated by Lecompte, and projects over the southwesterly-dipping flank beds on the exposed segment south of the railroad cut, from which point it follows them conformably in bedded form the entire length of the slope, as far down as exposed. Examination of polished and etched surfaces revealed, in a number of cases, that the generally observable, closely spaced, thin laminae which make up the ribbons are separated by short, closely spaced, vertical, pillar-like structures. Although there is thus no question as to their organic origin, it is not as yet certain whether these structures are stromatoporoids of the *Clathrodictyon* type or algae of the *Solenopora* type which construct frameworks in certain Niagaran reefs of Gotland (Hadding, 1941).

In the central-core area, the skeletal remains of these as yet undetermined builders form solid, heavy-built frameworks which may occupy as much as 80 per cent of the vertical surface exposure (fig. 3, *A*). Enclosed in these occur irregularly shaped lenses of fairly argillaceous dolomitic sediments, with little visible skeletal debris and made up principally of scattered, dismembered crinoidal remains belonging to two fragile, minute *Pisocrinus* species. In the projection of the core over the southwestern flank beds, the frame-builders tend to form solid, irregularly wavy, but essentially horizontally oriented, masses that simulate bedding when weathered. Numerous coarse-grained skeletal remains of atrypoid brachiopods with both valves intact, cephali and pygidia of *Sphaerexochus romingeri* and of *Encrinurus* sp. are found cemented into the framework by overgrowth of the frame-builders. Sediments which occur interlayered and interlensed with the frame-builders are slightly argillaceous, although considerably less than those in the central-core area, and they also contain a large percentage of fossil debris, which, on an average, is coarser in grain size than that in the central-core area. Larger, slightly more robust crinoidal skeletal elements, including *Pisocrinus campana* and *Gissocrinus quadratus*, are common; small col-

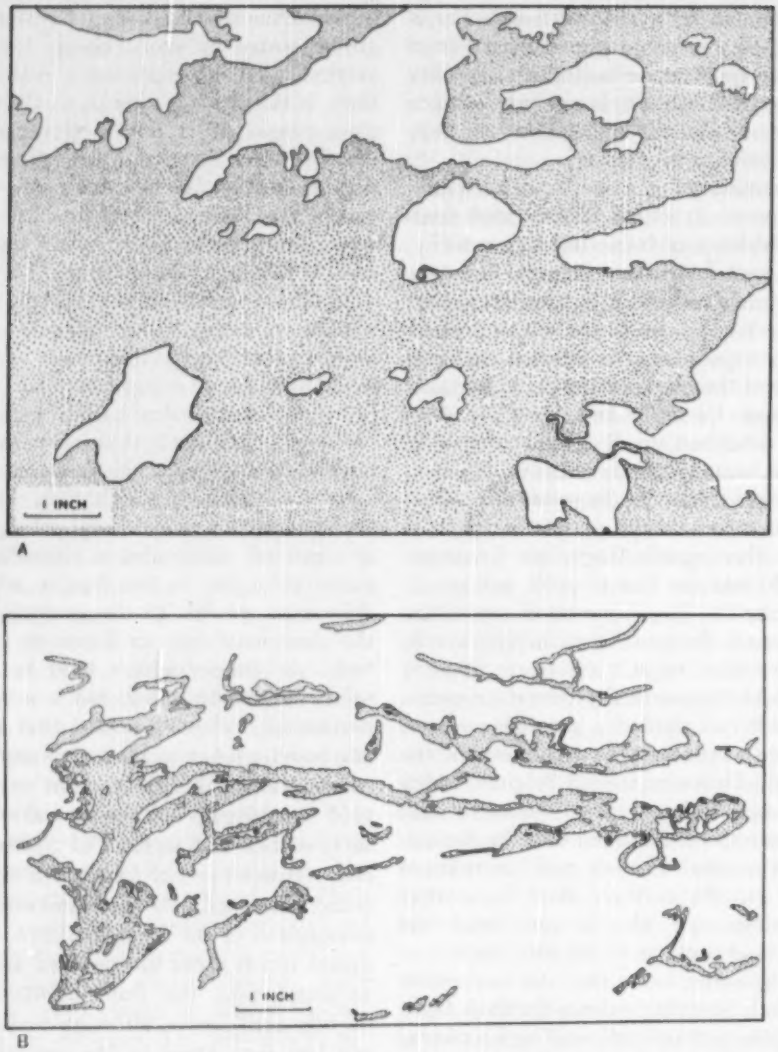
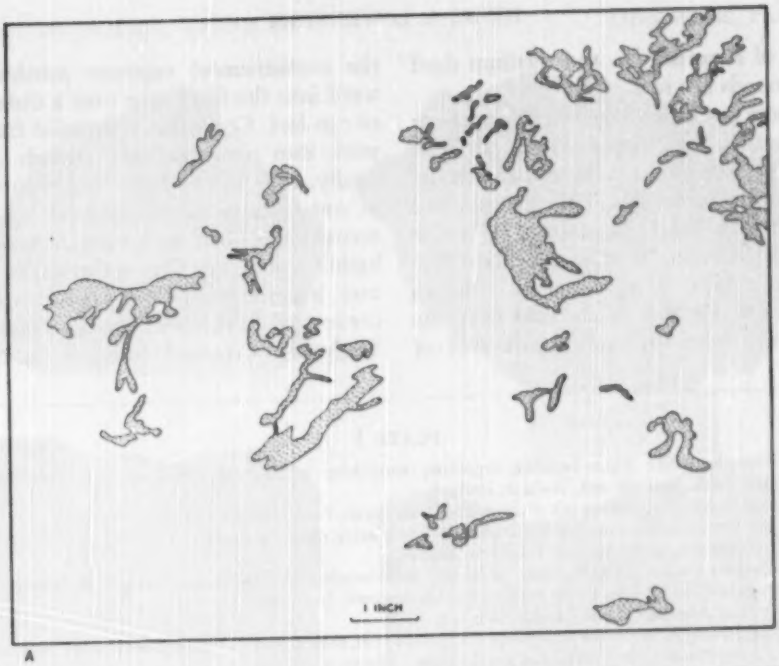
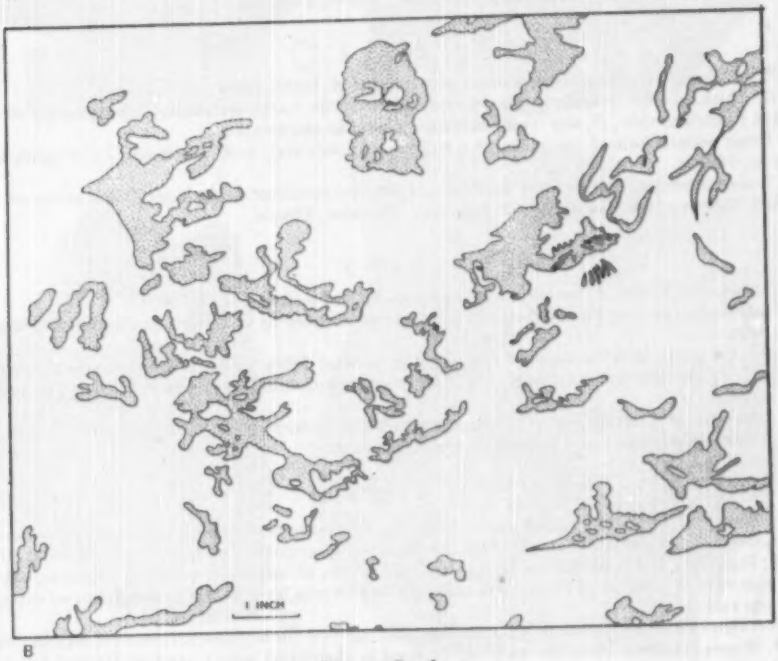


FIG. 3.—Growth characteristics of *Stromatactis*-like frame-builders at Wabash reef, Indiana. Actual tracings taken at intervals of 75 feet between 3, *A*, 3, *B*, and 4, *A*, and 10 feet between 4, *A*, and 4, *B*, from the central reef-core area southwestward into the reef flank. *Stippled areas: Stromatactis*-like frame-building organisms; *blank areas:* interstitial sediment fill.



A



B

FIG. 4.—See fig. 3

onies of *Favosites* sp. and medium-sized gastropods are rare.

There is a general outward decrease in volume of the frame-builders per unit volume from the central-core area to its southward projection. This decrease and the growth habit, as observable in the vertical section, is shown in the four sketches (figs. 3, 4) based on tracings made on celluloid sheets held over the exposure from the central-core area of the northernmost exposure southwestward into the flank area over a distance of 150 feet. Gradation from solid framework into irregular, lacy strands and, finally, into patchy habit and reduction in compactness of the skeletal ribbons accompanies their flankward decrease in building activity. This numerical reduction is accompanied by an increase in content of terrigenous clastic sediment. The mode of frame-building in the thin-

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PLATE 1

A, *Stromatactis*-like frame-building organism, enmeshing argillaceous dolomitic sediments. Southwestern reef flank, Wabash reef, Wabash, Indiana.

B, More loosely knit framework of *Stromatactis*-like forms. Southwestern reef flank, Wabash reef.

C, Lacy frame-builders, enclosing crinoidal, debris-bearing dolomite deposits. Loose block near southern fore-reefs, Thornton south quarry, Thornton, Illinois.

D, *Stromatactis*-like frame-builders, enclosing fossil-bearing, laminated, argillaceous sediments. Reef core, Georgetown reef, about 1 mile northeast of Georgetown, Indiana.

E, Mode of stromatoporoid frame-building.

F, Enlargement of one of the stromatoporoid pillars of *E*, showing type of preservation in the dolomitized form. Fore-reef, west-central portion of south quarry, Thornton, Illinois.

PLATE 2

A–B, Mushroom-shaped stromatoporoid growth form. *A*, bottom view; *B*, top view.

C–D, Thick, sheety, irregularly shaped reef-builders, with warty morphology, encrusting *Pycnostylus* coralites. *C*, surface view; *D*, side view, showing *Pycnostylus* encrustation.

E, Drusy growth form of encrusting reef-builder. *A–E* fore-reef, west-central portion of south quarry, Thornton, Illinois.

F, Sediment-binding reef-builders, forming a sheety encrustation of bioclastic debris at top of a reef-flank bed. Southern reef flank, Thornton main reef, Thornton, Illinois.

PLATE 3

A–B, *Favosites* colony. *A*, bottom view, encrusted by reef-builders; *B*, top view.

C–D, Detritus-catching Favositidae, encrusted and surrounded by encrusting reef-builders. *C*, top view; *D*, side view.

E, Solitary coral, showing mode of cementing of skeletal debris into framework by encrusting reef-builders. *A–E*, Thornton quarry, Illinois; *A–D*, fore-reef, west-central portion of south quarry; *E*, from loose rubble block.

F, Aggregate of *Conchidium* sp. shells, cemented into substrate by encrusting reef-builders (*light areas*). Northeast flank beds of Georgetown reef, Indiana.

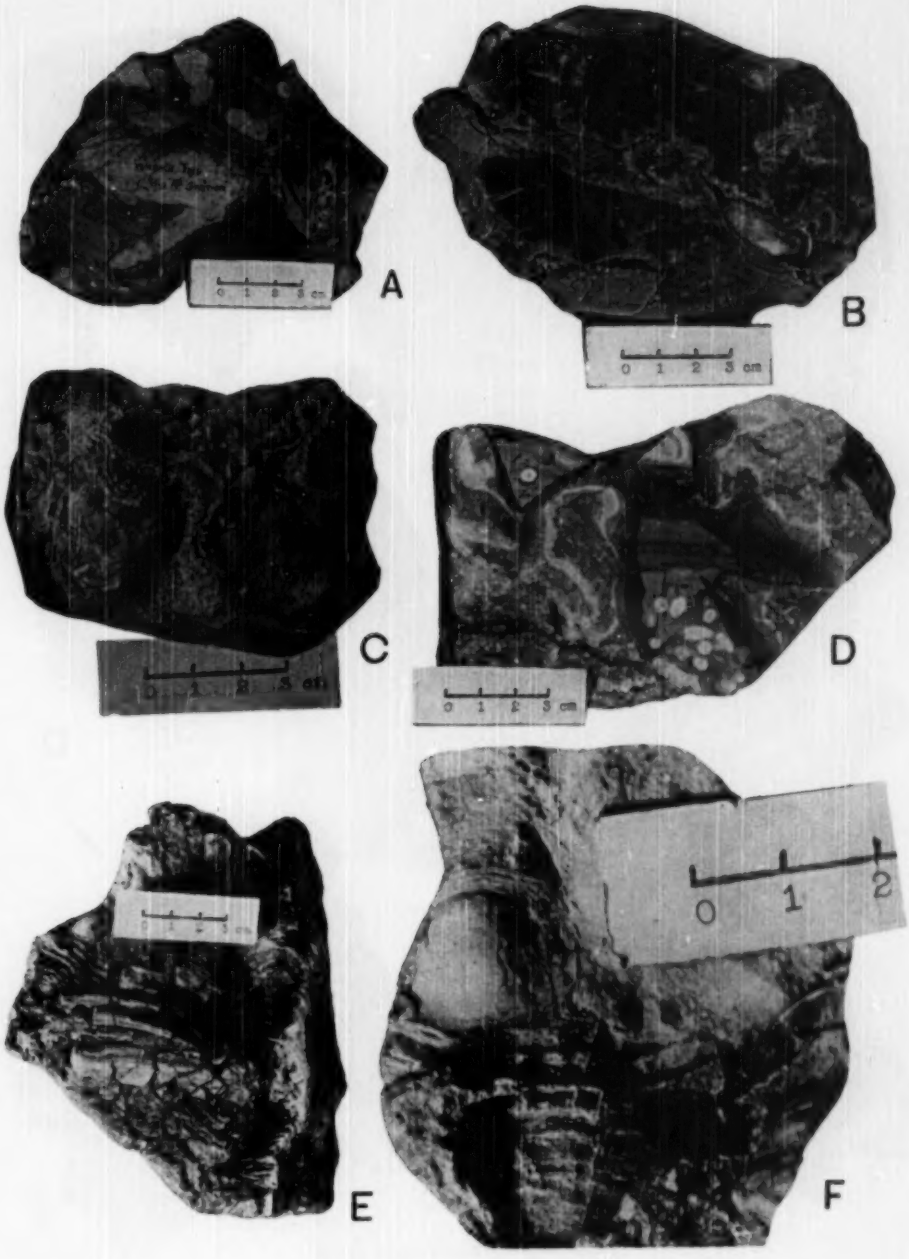
PLATE 4

A–C, Reef-flank beds composed of bioclastic debris, showing various degrees of sorting and preservation of debris constituents. Southwestern reef flank, McCook reef. Consumers Company quarry, McCook, Illinois.

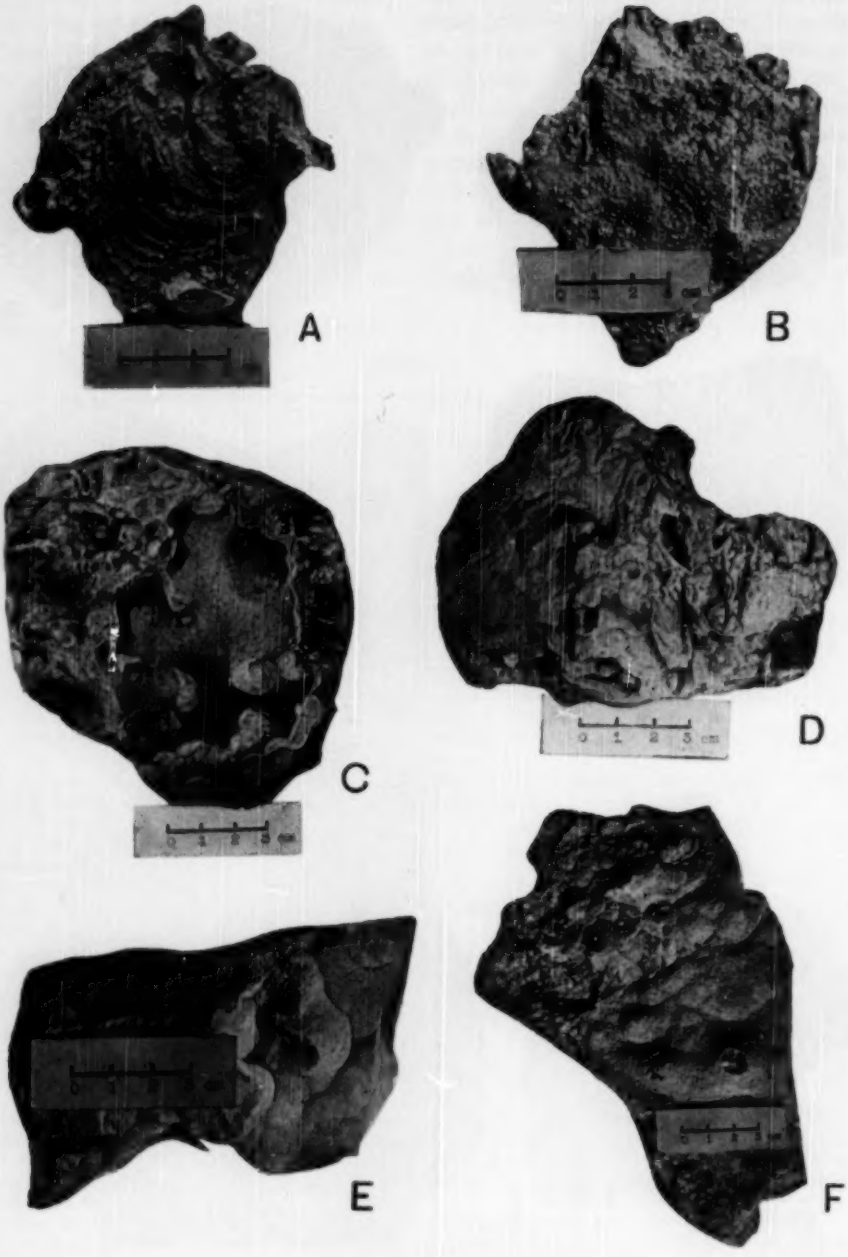
D–E, Reef-flank bed composed of encrusting reef-builders at bottom, overlain by bioclastic debris, as seen in side view, *E*, and capped by a more sparingly fossiliferous layer of argillaceous interreef deposits, as seen in top view of bed, *D*.

F, Top view of another reef-flank bed with bioclastic debris in the argillaceous interreef mantling film.

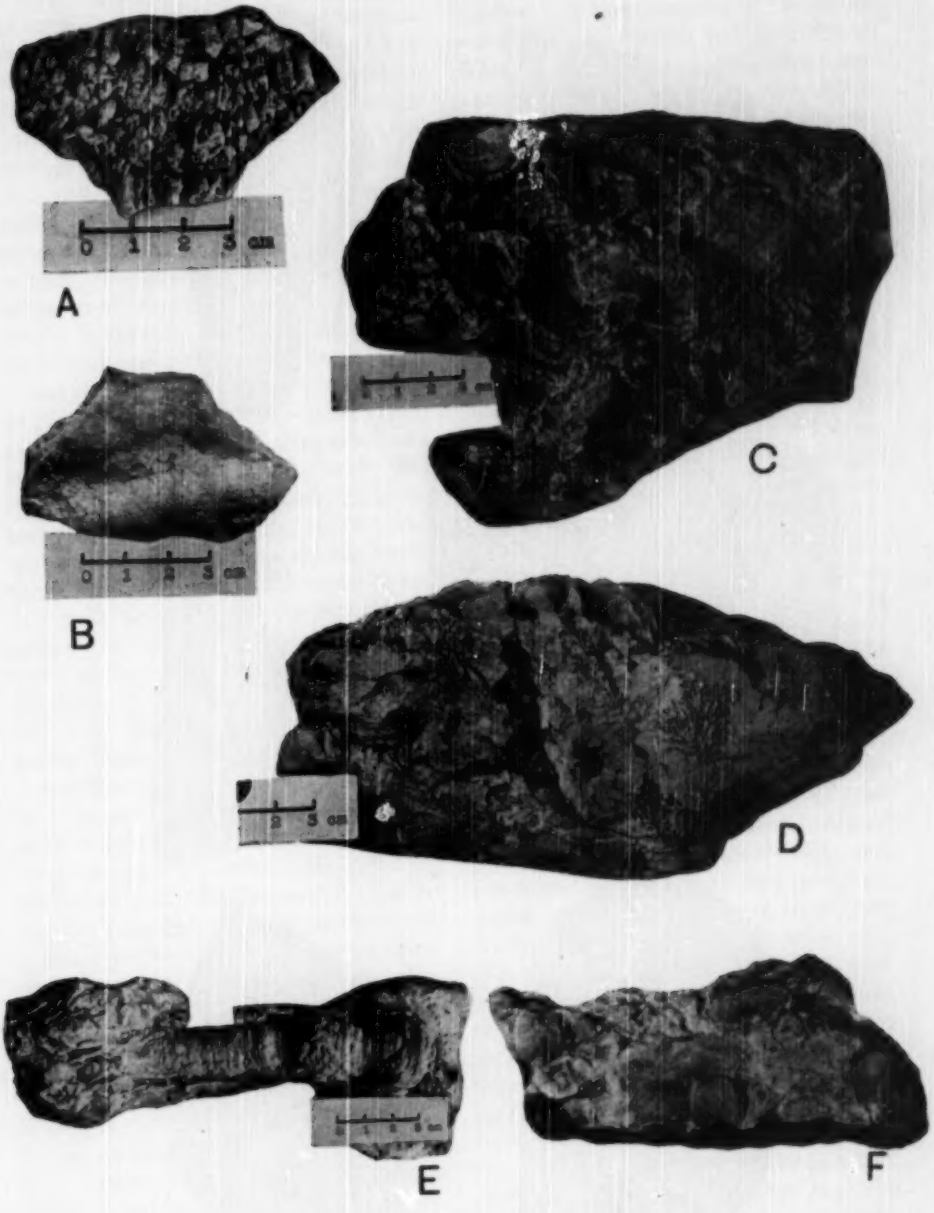
D–F, Western flank of Thornton main reef exposed in abandoned quarry, west on Halsted Street, near Thornton, Illinois.



Frame-forming reef-builders



Growth forms and modes of encrustation of reef-builders



Modes of encrustation of reef-builders

Reef flank detritus of bioclastic debris

bedded flank bed that extends southwest from the southern projection of the core differs in that a lenticular framework is maintained all the way down the flank as far as it is exposed (pl. 1, *A*, *B*). Bedding-plane outcrops here show that the framework encloses argillaceous lenses up to 2 feet long, representing former pools. Little fossil debris is, in general, visible in these lenses. Exceptions include a few minute, dismembered crinoidal skeletal elements of fragile build and a very few fragile bryozoans, medium-sized gastropods, and small, solitary corals.

Large, clear calcite patches occur both in the enclosed lenses and cutting across or locally replacing the frame-builders (Cumings and Shrock, 1928). On the outcrop surface, fossils, principally crinoids, are replaced by similar clear calcite. The lack of visible sutures, as contrasted with the prevalent cast and mold preservation in freshly broken rock and the general appearance suggest recent deposition from surface waters. Determination of the ratio of O_{16} and O_{18} of the calcite by Urey firmly established the suspected fresh-water nature of the calcite deposition. The light isotopic abundance of -9.4 per cent precludes precipitation of the calcite from marine waters, since a temperature of the order of about 68° C., computed on the marine scale, is prohibitive from the biologic point of view, but, if based on the fresh-water scale, a value of the order of 30° is quite reasonable. The cavities and filling are therefore recent and diagenetic.

The Wabash reef firmly establishes the existence of frame-building organisms, although their biologic affinities remain to be determined. These organisms exhibit further sediment-binding qualities by their sediment enclosures and by cementation of loose skeletal debris into their framework. Their microstructure, where preserved, indicates the frame-building and consolidation of the structure by a single species. As Cumings and Shrock (1928) pointed out, there are practically no corals. The few small *Favosites* colonies found by the writer confirm their view and show that the corals did not contribute to the building of the structure. Although the bryozoans might be regarded as accessories, functioning as mud-catchers, their numbers are so small that they could not have performed this task effectively.

It is important to note that rigid frame-building may extend from the massive core in bedded form into the flank element. Such a core type of frame-building may occur uniformly throughout entire flank beds and even a succession of them, as can be seen at the southwestern flank of the structure, or may occur only in the proximal portion of the flank beds adjacent to the core, as found in the northernmost exposure. In the latter there is perfect gradation from the highest, core-type portion down-dip into the clastic phase of the enclosing Mississinewa shale, the transition being marked by the thinning-out and, finally, the disappearance of the reef-builders and an intermediate clastic ratio of the lithologies (fig. 5). Organic frame-building and sediment-binding activity thus spread recurrently from the core area to its bordering flanks, supporting, in these specific cases, Lecompte's view (1938) that these flank portions reflect adjustment to environmental fluctuations to allow the spread of the organisms. The nature of the environmental conditions responsible for these phenomena will be discussed in connection with the origin of reef detritus. The significant point is the fact that reef-building activity extended past the core into the bedded, inclined flank beds, the only noticeable difference being that

the growth was interrupted intermittently, as indicated by the bedding planes, whereas, in the core, growth proceeded without interruption. The two are thus genetically identical, demonstrating that the flank element may be in part genetically core. The implications of this will be considered in the next section.

Another example of extensive frame-building was observed by the writer in

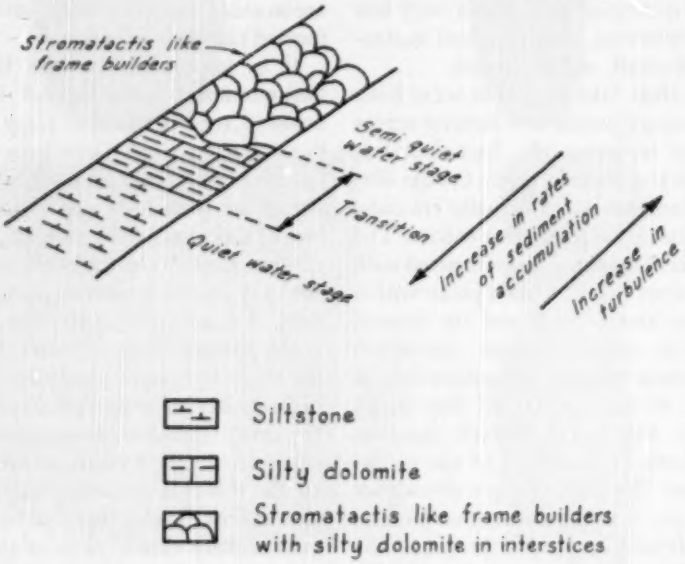


FIG. 5.—Diagram illustrating development of composite reef-flank beds of the Wabash reef, Indiana

the Georgetown structure (Cumings and Shrock, 1928), located about 1 mile northeast of Georgetown, Indiana. The Wabash River bisects the northern margin of the ridge-shaped structure, exposing northward-dipping, heavy-bedded flank deposits in the northern bluff and a massive core with some obscure flank beds on the southern bluff. The exposures show that frame-building organisms of the same type found in the Wabash structure are abundant. However, their density is considerably lower and the framework generally less compact and somewhat more irregular, branching, and discontinuous. The laminar structure with regularly spaced pore layers is here better preserved. The surfaces of the laminations appear to be perfectly smooth, strongly favoring algal, rather than stromatoporoid, affinities of the builders. The matrix around and enclosed by the frame-builders is fine-grained, dense, structureless dolomite with less than 5 per cent terrigenous clastic content (Cumings and Shrock, 1928, record about 2 per cent insoluble residue). Locally, however, in the highest preserved portion of the structure, the frame-builders enclose pockets of well-laminated sediments with a considerably higher proportion of terrigenous clastics (pl. 1, *D*). Fossil debris, except for a rare trilobite cephalon or pygidium, principally of *S. romingeri*, and a few crinoidal stem ossicles, is rare in the matrix of the

lower central-core area but increases conspicuously toward the ill-defined edge of the eastern flank. Tabulate corals, *Halysites* and *Favosites*, and solitary corals are present, but, together with bryozoans, they form less than 10 per cent of the total assemblage, the bulk being composed of a variety of small brachiopods and trilobites. Locally, near the top of the exposure, crinoidal debris becomes common, in places forming coquina patches.

The fossil assemblage changes conspicuously in the flank beds exposed in the opposite northern bluff. The frame-builders are prominent in certain beds of the lower 10 feet of the central flank section examined. They are generally explanate, irregularly sheety to coarsely meshy in form, and display debris-binding qualities. Brachiopods, principally heavy-shelled, large, robust *Conchidium* forms, are found in great number, firmly cemented into their framework (pl. 3, *F*). *Favosites* colonies are slightly more abundant and larger, on an average, than in the core section on the south side of the river. Also, a few colonies of *Romingerella* were observed associated with the tabulates, partially encrusted by the frame-building forms. A small number of bottom-attached corals thus aided in consolidating the flank. Their slightly greater abundance here as compared with the core area points toward zoning, which is clearly indicated by the abundance of the large *Conchidium* forms on the periphery, which are conspicuously lacking in the core area. Zoning is also shown by a shift in composition and size of the trilobite remains, whereas the size increase, coupled with a decrease in number of species, in the peripheral portion might be due to sorting on the basis of weights and shapes. The *Conchidium* zone, however, is real, as an occasional individual would otherwise have been trapped in the cavernous framework of the principal builders. Exposures do not, however, permit determination of whether or not the flank beds observed are the time equivalents of the observed core. In general, the Georgetown reef has the same problematic forms—the principal frame-builders and cementing agents, aided in holding down the unconsolidated bottom segments by the subordinate colonial coralline element—as does the Wabash structure.

Similar lacy frame-building on a subordinate scale has been observed in limited exposures of the southern flank beds of the Stony Island structure on the South Side of Chicago and in loose blocks near a series of fore-reefs skirting the southeastern periphery of the main Thornton structure south of Chicago (pl. 1, *C*).

At Thornton, reef-building activity is clearly demonstrated. Compared with the previous example, a considerably greater variety of building elements and a more complex building relation is displayed here. The Material Service Company quarry has exposed most of the eastern flank half of the main Thornton reef. In the abandoned southern quarry there is a considerable number of fore-reefs (Fenton, 1931; Bretz, 1939; Lowenstam, 1949*a*). Only the prominent, steeply dipping flank beds of the main reef are to be seen, except possibly in the central area of the west wall of the northern quarry, from which the northward- and southward-dipping well-defined flanking beds diverge. Here bedding is more massive and in part even obscure, as compared with the bordering, typical flank beds, and may consist in part of tongues of reef core interfingering with flank beds. Irregularly shaped, thinly laminated masses of reef-builders of unknown af-

finities are commonly traceable in the pure dolomite rock. Locally, these show warty to botryoidal surfaces, surrounding and, in part, encrusting colonies of *Favosites*, *Heliolites*, crinoid calyces, and brachiopods. The fairly coarse-grained dolomite that makes up these structures reveals only thin laminae arranged concentrically in the warty portions. In their gross morphology these forms closely resemble present-day lithothamnioids. A great variety of colonial corals, principally massive tabulates and Heliolitidae as well as colonial tetracorals, occur here and in the bordering flanks, becoming relatively most abundant numerically in the northern flank portion. Large colonies of slender *Pycnostylus* forms, some densely crowded over a square foot in area, occur in places. The corals, mainly low-lying forms, aided greatly in holding down the vast bioclastic debris composed of crinoids, brachiopods, bryozoans, pelecypods, gastropods, and trilobite remains, whereas the *Pycnostylus* colonies acted principally as detritus-catchers. There is no evidence of frame-building activity in this core area or peripheral-core portion or in the proximal flank-bed area. There is little evidence of detritus-binding activity except in the central portions, unless the structureless nature of much of the dolomite here has obscured the evidence and thus distorted the original proportional representation of the detritus-binders.

In general, the flank beds contain, on an average, a small percentage of tabulates, Heliolitidae, and stromatoporoids. The forms generally occur embedded sideways or in an upside-down position, showing burial in nonliving positions. They may have been derived, at least in part, from the eroded portion of the reef core which lies concealed to the west. Evidence of sediment-binding activity is rarely observable on the flank portion but has been seen locally on the southern flank and also on the western flank, which is exposed in an abandoned quarry west of Halsted Street, the location of which is shown on Bretz's figure 48 (1939). In the abandoned southern half of the operating quarry, a thin, warty to irregularly wrinkled sheet of thinly laminated organisms of unknown affinities, capping and binding bioclastic debris, may be seen, forming the top surface of a reef flank bed (pl. 2, *F*). Occurrence of the sediment-binders marks an interval of nondeposition, due either to by-passing or to local nonsupply of debris to the flank segment. This throws some light on the origin of bedding planes in the flank segment. Additional evidence is afforded by an identical wrinkled sheet at the bottom of a flank bed overlain by bioclastic debris made up principally of crinoidal remains (pl. 4, *E*). Several flank beds in the same quarry consist largely of a stromatoporoid and tabulate coral conglomerate, most prominently displayed in a more lenticular bed of considerably greater thickness than those enclosing it. A number of the stromatoporoids at the surface of this bed show small, sheety, irregular expansions, indicating that, following their being dumped on the flank, portions remained alive and continued temporary growth and encrustation of neighboring dead colonies, thus facilitating localized cementation of the flank debris.

Extensive building activity was noted in a small fore-reef in the west-central portion of the abandoned southern quarry. This fore-reef is located near the southeastern limit of the flank of the main Thornton structure. Its massive core displays a wide range of building modes. Some stromatoporoids, in tall, columnar form, are seen here, rising as

much as a foot vertically from an encrusting sheety base. Some enclose larger sediment segments by lateral expansion from column to column (pl. 1, *E*, *F*). Thick, irregular, sheety encrustation with warty or irregular surfaces are common, enclosing slender *Pycnostylus* coralites and encrusting *Favosites* colonies (pls. 2, *C–E;* 3, *A*, *B*, *E*), and there are illustrations of the mode of frame-building by means of cemented detritus-catchers and sheety cementation, as seen in plate 3, *C*, *D*, where a colony of massive branching Favositidae, acting as detritus-catchers, is partially cemented over and in part enclosed by a sheet of stromatoporoids, firmly cementing the detritus between the corals onto the framework. In several cases it has been possible, despite dolomitization, to identify, on the basis of typical astrorhizae, the cellular, sheety encrusting forms as belonging to the stromatoporoid genus *Clathrodictyon*. Other forms of this type lack surface preservation, and there are some which appear to belong to different genera, if the widely spaced circular markings are actual surface features. Low, mushroom-shaped stromatoporoid forms with narrow bases of attachment seem to mark protected topographic positions with low rates of deposition and space for lateral development (pl. 2, *A*, *B*). The prevalent nodular to semiglobular shapes of the silicified stromatoporoids, which occur abundantly in the comparatively low clastic interreef beds near the top of the east wall of the same quarry just south of the incline, are generally missing in the reef. Well-preserved individuals of both morphologic types will be required to determine whether we are dealing with habitat-differentiated species or environment-controlled growth forms. In either case, differentiation of morphologic types as related to environment is definitely observable, a point which Lecompte (1938) categorically denied.

The gross expansion pattern of frame-building activity in the cores of the Wabash, Georgetown, and subsidiary Thornton structure, although irregular and more patchy in the last two, is most pronounced in the vertical dimension, whereas in the flanks of the first two it is principally parallel to the slope. The low bank-type development of building activity, characterized by horizontally bedded stromatoporoid growth outside the structures, as found in Gotland (Hadding, 1941), is generally lacking in the area under consideration. Only in a single case have we observed horizontally bedded, massively laminated structures in the north face of the Consumer's quarry at McCook, west of Chicago. This quarry exposes the southern flank portion of the large McCook structure (Bretz, 1939). Recent relocation of the incline along the north wall has exposed the contact of the base of the inclined flanking beds with the underlying horizontally bedded interreef deposits. The topmost portion of the latter consists of thinly, and in part minutely, laminated ribbons of pure dolomite of the type characteristic of the frame-builders noted above; evidently, it is a peripheral bank-like expression of building activity which was terminated by burial beneath the flank front, which advanced southward from the main structure.

Bryozoa have been commonly described as Niagaran reef-builders. In the reefs examined, they were fragile, commonly branching, and less commonly fan-shaped fenestelloid types. Their rigid calcareous skeletons evidently lacked the flexibility which enables the present-day Gorgoniacea, among the corals, with similar building plans but organic skele-

tons, to inhabit fairly agitated reef portions. The Niagaran bryozoans were not, therefore, rough-water dwellers. This is fully borne out by the burial distribution. In the frame-building portions of the structures they are numerically conspicuous only in bioclastic debris accumulations, which apparently filled more protected quiet to semiquiet depressions, crevices, and pools in the midst of the topographically raised centers of building activity. Their greatest abundance, however, is found in certain flank beds composed entirely of bioclastic debris, and, wherever interreef deposits with high terrigenous clastic content intertongue with the latter, they are found generally concentrated in the transition zone. They can thus hardly be regarded as reef-builders but were potential agents of sediment retention in the quiet to semiquiet waters. Even here, however, their role in such functions must have been minor or rather temporary, as their generally sidewise-embedded burial positions show.

In summary, we may state that there is ample evidence of frame-building, sediment-binding, and detritus-catching elements, thus establishing the presence of organisms capable of erecting rigid structures and potentially capable of functioning as more resistant elements. Both tasks were repeatedly performed by one and the same builder. There is still uncertainty as yet as to the biologic affinities of a large segment of the critical builders, but the biologic nature of their skeletal remains is beyond question.

SOURCE ELEMENTS OF BIOCLASTIC DEBRIS

Although reef-builders are organisms essential to the formation of present-day reefs, it has been shown that they commonly form a subordinate part of the whole structure. The bulk consists of bioclastic debris (Walther, 1893; Ladd, 1930). The debris, the so-called "coral sand," consists, moreover, of only a subordinate to insignificant proportion of the remains of the reef-builders. The bulk of it is derived from reef-dwelling organisms. The bioclastic debris fills the interstices and the lagoons, if any are present, and materially aids in building up the reef from relatively deeper waters. It makes possible the horizontal expansion of wave-resistant reefs, provided that it is held down and cemented by organic and possibly inorganic agents in the wave-exposed portions of the reef. Debris contribution is most effective on reefs growing in shallow-shelf seas, particularly where the bottoms are located near effective wave base. Rapid peripheral accumulation raises the bordering bottoms into the zone of intense building activity and thus hastens reef expansion and stimulates the formation of subsidiary centers of growth. In contrast, the debris from oceanic reefs with narrow, deep-lying platforms does not perform such functions. The Niagaran structures belong to the shallow-sea, bottom-type reef.

There is ample evidence that the Niagaran structures had potential sources of bioclastic debris. Large, robust fossils characterize the flanking beds and even the highly altered, dolomitized core, in sharp contrast to the enclosing deposits that contain, on the average, a lower fossil content marked by small size and fragile build. The bioclastic debris consists of a highly varied assemblage of crinoids, pelecypods, gastropods, cephalopods, brachiopods, and trilobites. The outcropping Elmhurst, Thornton, and McCook structures and the impressive 500–1,000-foot-thick structures in southern Illinois and Indiana (Lowenstam,

1948*b*, 1949*a*) all show that crinoids constitute the most important source of debris. The laterally jointed, flexible, fishnet type of arm structures of the endemic *Crotalocrinites* and the reclined, marginally ending, anal-tube structure of *Siphonocrinus* are examples of specific adaptation not found off the reef structures. The forms with greatly shortened stems and stereom-cemented roots, known only in *Euspirocrinus spiralis* (Bather, 1893) from Gotland, Sweden, that, beginning in the Devonian, became a standard design in rough-water reef crinoids, have not as yet been observed on the Niagaran structures on this side of the Atlantic. The crinoids, forming enormous colonies, grew principally in habitats with freely circulating waters. These were, however, apparently shielded from continuous high turbulences save for the occasional bottom-sweeping storm waves that produced the sorted crinoidal coquinas of many reef flanks. Their skeletons that disintegrated post mortem into skeletal sands clearly exclude the crinoids as potential reef-builders but make them an ideal source of bioclastic debris. They appear, throughout the Paleozoic and most of the Mesozoic, to have occupied the same ecological niche and performed the same important detritus-supplying function that the jointed calcareous algae, *Halimeda*, do today. Some representatives of the other phyla, although greatly subordinate to the crinoids as detritus contributors, form the main constituents of the debris in small areas. In rare cases they may become the dominant contributors, as on the Georgetown structure, where *Conchidium* provides the bulk of the detritus.

Thus, in addition to the evidence of frame-building and sediment-binding action by organisms, it may be concluded that there was, as there is in the present-day reefs, an adequate source of bioclastic debris available for similar functions, although it was composed of quite different habitat dwellers.

ECOLOGIC GROSS FEATURES OF THE ASSEMBLAGES AND THEIR BEARING ON REEF GROWTH

If the structures in question are true reefs, we would expect to find fossil assemblages which are indices of rough- and semirough-water habitats. Cumings and Shrock (1928) were the first to point out the existence of a general sharp differentiation between the assemblages on and off the structures. The writer has substantiated this observation in his study of the interreef assemblages and has shown that the Racine and Guelph faunas are confined to the structures, actually forming their characteristic dwelling elements (Lowenstam, 1948*a*). Subsequent field studies in northwestern Illinois, Wisconsin, and Ontario have fully borne out this thesis. Wherever the structures are enclosed in deposits of rough-water origin and lack discernible flank deposits, as in the clastic-free belt (Lowenstam, 1949*b*), the Racine and Guelph assemblages are found concentrated mostly near the structures from which they had been swept off after death. The contrast in assemblages is most conspicuous where sediments rich in terrigenous clastics border or enclose the structure and is most impressive where there is interfingering of the two, as at Thornton and Elmhurst in the Chicago area (Lowenstam, 1948*a*). In general, the assemblages on the structures show considerably greater fossil density and, where the preservation of the original textures is clearest, a considerably greater number of species, mainly endemic forms, larger and more robust than those in the surrounding deposits. The

rough-water interreef assemblages have a transitional character.

The assemblages of builders and associated dwellers show considerable variation from place to place on individual structures. The most conspicuous differences are those between the flank beds and the cores. Also, there is marked contrast of the assemblages between structures.

Lecompte (1938) called attention to the contrast between core and flank assemblages in the Bluffton structure in northern Indiana and emphasized the greater abundance of corals on the flank as compared with the core. This peculiarity formed part of Lecompte's argument that these structures were not true reefs.

This type of comparison of the flank and core elements is fallacious because the base of the flank-bearing structures is, as a rule, not exposed and the top has been removed by pre-Pleistocene erosion. Only random sections, the positions of which are not known in relation to the total structure, have been studied. Truncation of most of the flank beds by the erosion surface clearly shows that these inclined strata must be correlated with core sections, now removed, which were younger than the portions preserved. Even where the lowermost flank beds seem to merge into the core, as noted by Lecompte in the Bluffton example, there is no assurance that the preserved portion of the core corresponds in age to them. The flank beds could have been formed much later than the core against which they lean if the structure rose initially from quiet water below effective wave base. The structure would remain steep-walled, essentially free of bordering debris, until it was raised above wave base. Then the sweeping and grinding action of the turbulent surface water would affect it. Debris carried off the structure would accumulate at its foot and would therefore be considerably younger than the core against which it was deposited. The skeletal remains in this flank deposit would certainly differ from those of the bordering core, inasmuch as they were derived from an assemblage characteristic of the wave-resistant phase, whereas the bordering and older core assemblages were those of the quiet-water phase. Larger coral colonies in the flank beds, therefore, would not seem anomalous.

This argument is by no means theoretical, as it can be shown that the bottoms of certain Niagaran structures in the low-clastic belt remained below wave base even after the structures attained wave resistance. The erosion-scalped Elmhurst structure west of Chicago (Bretz, 1939; Lowenstam, 1948*a*) bears out both points. Although the core has been removed by quarrying, recent clearing of the rubble from the quarry floor, where the core once was, has exposed a number of unusually large *Syringopora* colonies, which were the pioneer population of the structure. A typical quiet-water assemblage established in the surrounding argillaceous deposits shows that the structure started in waters below wave base. The next 42 feet of interreef strata, which were laid down while the structure rose and began to accumulate peripheral flank beds, consists of similar quiet-water deposits. The intertonguing on a small scale with the flank beds close to the top of this unit establishes partial time equivalency. Pure dolomite beds dominate the top 19 feet of the interreef section. These can be traced laterally into the erosion-scalped flank beds, thus demonstrating a contribution of fine carbonate from the portion of the structure which had been eroded prior to quarrying operations to the bordering

bottoms. One critical detail, namely, the contact relations of the lowest interreef strata with the reef core, cannot be checked because of removal of the core and concealment of the lowest flank beds by the quarry incline.

Much of the main structure at Thornton can also be shown to have grown while the surrounding bottoms were below wave base. Here the unusually extensive flank beds intertongue with muddy, quiet-water interreef deposits (Lowenstam, 1948*a*). The recently filled-in quarry, about a block east of Cottage Grove Avenue at about Ninety-second Street, which exposed westward-dipping flank beds of the Stony Island structure (Bretz, 1939), showed analogous intertonguing of the flank deposits with argillaceous, quiet-water interreef beds.

The history of the Elmhurst, Thornton, and Stony Island structures, shown by the quiet-water character of the surrounding and intertonguing deposits and their enclosing fossil assemblages, implies a gradual, although ultimately radical, vertical change in the composition of the fossil assemblages in the structures. The most pronounced contrast in assemblage composition would occur between the initial quiet-water and the ultimate rough-water habitats. In the quiet-water phase the environmental conditions would be essentially identical on the growing structures and their surrounding bottoms, as the fine muds carried in suspension by the sea could settle freely on both. The frame-builders would tend to grow mainly in the vertical dimension, to avoid suffiocation by mud coverage. The niche dwellers would be primarily the same as those populating the surrounding bottoms, save for a possible selection of hard-bottom dwellers. Thus there should be basically little difference between the organisms on and off the structure, with the exception of the building elements.

While passing through the interface of effective wave base, expansion in the horizontal dimensions of the frame-builders should prevail as a result of decrease in quantity and more intermittent settling of suspended sediment. A denser and more varied population of niche dwellers is also to be expected. The reef dwellers would now differ from the population of the surrounding quiet-water bottoms. In the final rough-water, wave-resistant stage the trend toward intense building and consolidation through debris-binding activity would reach its climax. Reef-building would be mainly in the horizontal direction, provided that sea level remained stationary. The optimum favorable environment created by the builders, that is, shallow, agitated, well-lighted, and well-oxygenated waters in combination with a complex surface relief, would provide surface and shelter for an abundant and varied fauna, totally different from that of the initial quiet-water phase.

None of the three structures noted above show vertical changes required by this oversimplified pattern of evolution, owing primarily to inadequate vertical, and also unsatisfactory horizontal, exposures, which are limited to parts of the flank deposits and to a possible exposure of the edge of the core of the Thornton reef and partial sections of a number of small fore-reef cores on its southern and southeastern periphery. Whether or not the flank beds constitute talus fans, principally of reef outwash, as maintained by all who consider the structures to be reefs, or remains of organisms which died in place, as implied by Lecompte, there are noticeable gross differences in fossil composition between the Thornton and the Stony Island flank beds, on the one

hand, and those of the Elmhurst structure, on the other. The Thornton and Stony Island assemblages, as a whole, consist of a variety of corals, stromatoporoids, the unknown encrusting organisms, brachiopods, pelecypods, gastropods, cephalopods, and trilobites. In only a few cases are the crinoids dominant. The varietal abundance, the preponderance of endemic species and robust forms, more pronounced at Thornton than at Stony Island, are consistent with the assemblage composition and character expected in the rough-water phase and with the stage of different ation attained in Niagaran biota.

By contrast, the Elmhurst assemblages are poor in varietal forms but rich in individuals. Crinoidal remains of large robust forms are generally prevalent and form seams of coquina, apparently derived from comparatively few species. Corals are common in places but, in general, form a subordinate element of the faunas. Few species are represented. Brachiopods, pelecypods, gastropods, and trilobites are rare. These assemblage characteristics are those to be expected at the intermediate stage, namely, just following the rise above effective wave base.

Local conditions, such as a shielded source area on the structure or, if we assume that the forms died in place, a shielded growth habitat, might be responsible for the peculiarities of the Elmhurst assemblage, so that the first alternative might be open to question.

The complete sequence of postulated vertical faunal changes has nowhere been observed to date. This is not surprising, in view of the shallowness and general inadequacy of exposure and the erosional truncation of the flank-bearing structures. The Wabash structure in northern Indiana is, however, an example that confirms the essence of the thesis, as it illustrates a definite vertical change in building behavior that is accompanied by shifts in composition of the niche dwellers. The exposed, eroded portion of the structure records an initial vertical building stage followed by a horizontal, southward-directed growth stage. This horizontal expansion is correlated with a tendency of the builders to assume bedded, low-lying forms and also with a marked increase in niche dwellers. The latter are largely newcomers, unlike those in the earlier stages of the structure, and they are also more abundant and slightly more varied in species. The Wabash fauna is poor in contrast to the rich assemblages of Thornton and Stony Island. The crinoids are, on an average, even more fragile in build than those of the Elmhurst flank beds. The building and cementing functions at Wabash are carried out only by the *Stromatactis*-like forms. In the late expansion stage a few small *Favosites* colonies participated slightly in the building activity.

Most unusual, however, is the makeup of the niche dwellers. The Pisocrinidae, brachiopods, bryozoans, and trilobites of the initial vertical building phase form an assemblage which characterizes the quiet waters of the muddy, interreef bottoms. This assemblage has been observed on the flanks of the Thornton and Elmhurst structures, but it occurs only in tongues of interreef sediments within the flank beds. In a single case it was found in interstromatoporoid cavities and small pockets at the northeastern fringe of the southern Wabash structure (Lowenstam, 1948*a*). These assemblages, together with the characteristically associated terrigenous clastics, normally mark protected, low-lying embayments and pools on the reef periphery, whereas at Wabash they are normal features of the initial

stage of the core. The lower portion of the core of the Wabash structure is thus a product of quiet waters and is therefore a clear-cut example of the postulated initial quiet-water stage.

In the period of horizontal expansion, the niche dwellers show affinities to both quiet and rough waters. *Pisocrinus campana* and *G. quadratus* are typical inhabitants of the quiet-water bottoms, whereas *S. romingeri* ranges from quiet- to rough-water habitats. The abundance and size of *S. romingeri* is consistent with its preference for semirough-water habitats, which are generally characterized by greater size and numbers of this form than are quiet-water deposits. The abundance of comparatively smooth atrypoid brachiopods and the first, although meager, appearance of gastropod types characterizing the final rough-water stage, together with the first appearance of *Favosites*, and, finally, the tendency of the building elements to show low-lying, massive encrusting growth establish a semirough-water habitat. A decrease in content of terrigenous clastic materials from that of the earlier stages further supports this conclusion.

In the Wabash structure the change of building behavior of the frame-constructing organisms and the concurrent changes in the niche-dwelling assemblage is what we would expect on a reef surface rising from the deep, quiet waters to effective wave base and semiturbulent waters. The intertonguing of erosion-decapitated flank beds, correlative with a higher, erosion-removed portion of the core, with normal Mississinewa siltstones shows that the habitat change from quiet to semirough water and the concurrent horizontal expansion of the reef were due to the raising of the building surface by the organisms themselves and not by passive shallowing of the sea. Likewise, the habitat changes cannot be related to a decrease in regional supply of terrigenous clastics.

The contrast between building behavior and niche dwellers of the flank and of the core of the Georgetown structure in western Indiana has been noted previously. This contrast is obscured by the Wabash River, which lies between the northern flank section and the core structure to the south. Whether the flank beds are synchronous or younger than the preserved core section is not clear. The faunal differences between the two could be interpreted as an expression of successive vertical growth stages, with the flank faunas characterizing an ecologically distinct fringing zone of a shallower-water phase. In any case the assemblage marks a stage in turbulence intermediate between that of the uppermost Wabash and that of the Stony Island and Thornton structures.

If the outcrop records are pieced together, a trend emerges, starting with the pioneer population of the Elmhurst structure, through the Wabash, the higher Elmhurst, the Georgetown, the Stony Island, and the Thornton reefs. This sequence shows that, in the low-clastic belt, frame-building and sediment-binding organisms were able to build discrete topographic structures which they raised from quiet waters below wave base into the agitated waters above wave base. It has also been shown that, correlative with the rise of the growth surfaces, the corals became conspicuous, both in species and in abundance, and functioned principally as a bottom-consolidating and sediment-retaining element. An impressive shift in character of the niche dwellers accompanies the gradually changing habitat conditions, beginning with an insignificant, species-poor, quiet-water, interreef fauna and ending with a

rich, highly varied, rough-water assemblage which greatly exceeds the building assemblage in volume.

Although denied by Lecompte, it can be demonstrated that there were vertical changes in faunal assemblages, changes which were controlled by environment. The frame-builders changed from dominantly narrow-based, vertical forms to horizontally spreading forms. Finally, sharp differentiation between the assemblages on and off the structure are most clearly shown where contrasting environments border each other. Branching Favositidae mark the protected habitats, whereas the massive hemispherical types are principally found in the unprotected areas of the turbulent-water stage, as observed on a small fore-reef at the southern edge of the flank of the main reef at Thornton. It is not certain whether any two of the massive and branching types are conspecific, that is, adaptive growth forms, or whether they are distinct species. Topographically controlled ecologic differentiation, in either case, can be shown to have existed in the Thornton fore-reef. It is conceivable, although yet to be demonstrated, that, at this early period of experimentation in reef-building, environment selection of preadapted types prevailed over specific morphologic adaptation to the environment, as has been demonstrated for present-day reef-building corals by Vaughan (1919). That the building elements comprise, for example, apparently few species and that crinoids, brachiopods, and trilobites are dominant among the niche dwellers is not disturbing, nor is it disturbing that we do not find boring echinoids, *Halimeda*, and lithothamnioids. In other words, the Niagaran reefs had totally different faunal assemblages from those of modern reefs—a point which impressed Lecompte as critical. It is the raw material available in the Silurian upon which selection operated in determining the character of the assemblages of the highly specialized habitats. The existing radical differences between Niagaran and present-day reef biota are, therefore, to be expected. We wish to learn, in terms of present-day mastery of the task, how successful the Niagaran assemblages were in achieving their goal and by what means they succeeded in doing so.

One critical point in Lecompte's argument is not yet disposed of, namely, the lack of ecologic zoning on the structure itself. From the ecologic point of view, we would expect biotic zoning at the wave-resistant stage. This problem must be properly explained in order to establish the reef character of the Niagaran structures. The flank beds were, at any given time, structurally lower than the reef surface and therefore at levels of lower turbulence. Zoning might occur on the flanks down-slope, but it would be less pronounced than on the reef surface. Both the cores and the flank beds of all outcropping flank-bearing reefs, such as Thornton, have been deeply truncated by pre-Pleistocene erosion. Therefore, the critical portion which would contain the record of zoning has little likelihood of preservation. As far as is known, the preserved flank beds are all younger than any accessible portion of the core. The detritus contributed by the surface community to the flank—to its lower portion, in particular—cannot reflect the surface zonation. Only a core and younger flank beds exposed together, with both recording and wave-resistant phase, would show zoning of the sessile, bottom-cemented core forms. No such exposures have yet been recognized, either because of large-scale alteration of fossil identity by dolomitization or because proper out-

crops have not been located. The most logical places to investigate the problem are the subsurface reefs in southern Illinois and southwestern Indiana, in which the surface topography of the reefs has been only slightly eroded. We do not have, nor are we likely to get, enough cores here to settle the question.

WAVE RESISTANCE

The Niagaran structures have been shown to be discrete topographic forms erected by organisms that were capable of raising them from quiet-water bottoms to wave base and above. But to call these structures true reefs, it is necessary to show that they attained wave resistance. Wave resistance implies that Niagaran reef-builders existed which were capable of coping with the turbulent surface waters and possibly even with intermittent intertidal exposure. The fossil record should show evidence of ecologic zoning and, as Lecompte (1938) demanded but failed to observe, evidence of marginal, ridgelike structures built by encrusting organisms comparable in function and effect to the lithothamnion ridge of present-day reefs. We also have failed, for reasons differing from those of Lecompte, to find either the zoning or the marginal ridge.

We must fall back, therefore, on the records of the effects of the physical forces which act upon present-day wave-resistant reefs. The full impact of the destructive forces constantly at work on present-day reefs finds its expression in the enormous quantities of bioclastic debris. The deflationary forces, principally the product of wave and surf action, are materially aided today by the reef-borers, comprising a host of forms that range from algae to crustaceans (Otter, 1937), which tunnel and surficially dig, perforate, and etch the dead portion of living builders and the framework, weakening and thus accelerating the destruction of the framework. The nibbling action of reef fishes and the destruction by crustaceans are contributing, although probably minor, deflationary factors. The detritus that fills in and spreads over large portions of the surface of the structure is, in many cases, built up to form islands which become fixed through cementation and anchored by vegetation. Thus the detritial material retained on the reef structure proper tends to convert it, at the climax of the wave-resistant stage, into a sand waste and ultimately restricts the building activity to narrow peripheral zones. The reef is then actively expanded outward only; and, if it is in the path of prevailing winds, it grows to the windward and is essentially static on the leeward margin. Estimates of the detrital cover range from three-fifths (Walther, 1893–94) to less than 1 per cent (cited in Adams, 1930) of the total surface. Large quantities of detritus are carried outward; the coarsest fraction accumulates largely on the slopes of the structures, the finest is carried outward and contributes to the sediments of the surrounding bottoms. The finest clay-sized fraction, the "reef milk," is carried farthest in suspension and has been observed as far as 12–14 km. from reefs (Agassiz, 1888).

Hall's reference (1862) to the bioclastic debris of the Wisconsin Niagaran reefs implies wave resistance. More clearly defined statements of Chamberlin (1877) and Grabau (1924), as well as Cumings and Shrock's (1928) emphasis on the detrital nature of the reef flank, led to general acceptance of the wave-resistant concept. Yet Roll (1934), in the analysis of the Upper Jurassic "sponge reefs" of southern Germany, voiced the opinion that the Niagaran reefs of North

America, like many others, were bank-type facies lenticles. Roll contended that these structures were genetically unrelated to reefs and consequently could not have been wave-resistant in the sense of true reefs. Reinterpreting Wiman's sketch (his fig. 16) showing the relation of the Silurian Gotland reefs to their surrounding strata, Roll concluded, from analogy to the Jurassic "sponge reefs," that the detritus accumulations fringing the lenticles were the products of intermittent bank deflation. Lecompte (1938), who briefly examined the Niagaran reefs, questioned the detrital nature of the flanking beds, which Cumings and Shrock had interpreted as talus fans, and contended that, although they are composed principally of bioclastic material, they are unlike present-day reef debris because the organic remains are fragile and there is lack of evidence of a beach-rock type of coarse outwash.

Subsurface evidence available in the last five years has provided us with a three-dimensional picture of the effect of these structures on their immediate surroundings and their area of distribution in general (Lowenstam, 1948*b*, 1949*a*). The recognition of a zone of turbulence around the Marine structure (fig. 6) leaves little doubt that wave resistance was attained in this case.

The origin of the flank beds, a matter of dispute, will be considered first. As pointed out above, only those flank beds which show lithologic and faunal evidence that they originated in the final growth stage will be considered. The flank elements of the outcropping McCook, Stony Island, and Thornton structures of northeastern Illinois and the buried Marine structure in southwestern Illinois are of this type. In each case the flank deposits are located in a region of terrigenous clastic sedimentation, so that we are provided with a means of determining whether we are dealing with intermittent storm-induced or continuous structure-induced turbulence. Except for well-defined tongues and laminae of terrigenous clastic-rich sediments, which constitute temporary overlaps of quiet-water interreef deposits, the flank beds at Thornton and Stony Island consist of high-purity dolomite. At Marine they are composed of dolomitic limestones and pure limestones. The typical flank materials forming either the entire flank deposit, as at Marine and McCook, or the bulk of it by volume were thus laid down under continued turbulence. The Marine flank deposits, in their undolomitized form, consist of a coquina of bioclastic debris with clear calcite commonly filling the interskeletal spaces. Crinoidal remains are prevalent. The dolomitized flank beds at McCook, Stony Island, and Thornton contain numerous individual beds consisting solely of debris. Others contain streaks of debris. Both types may contain intricate patches of dense dolomite. Crinoidal debris is prevalent and forms graded coquinas, with varying grain sizes from coquina layer to coquina layer. At both Thornton and Stony Island, beds occur made up entirely or prevalently of coarse stromatoporoid and coral debris. A large portion of a flank bed may consist principally of large, heavy-shelled, single-valved *Conchidium* shells, as has been observed at Thornton, or of fragile bryozoans, as seen at McCook (pl., 4, *A*). Beds of unsorted debris made up of a great variety of forms are common, particularly at Thornton, chiefly near the concealed core. The best-preserved debris observed to date occurs in certain layers on the southwestern flank of the McCook structure in the quarry of the Consumers

Company (pl. 4, *A–C*). Fossil preservation is considerably better at the contacts of shaly laminae or tongues and layers of bioclastic debris, as seen at Thornton (pl, 4, *D*, *F*). The contact is marked by a sharp decrease and, as a rule, a change in composition of fossil remains. It is not clear whether the dense, fine-grained dolomite beds, layers, and patches are texturally obliterated, sand-sized debris or fossil aggregates. Complete obliteration of fossil character by dolomitization has been observed on a small scale at Marine (Lowenstam, 1948*b*).

The foregoing observations show clear-

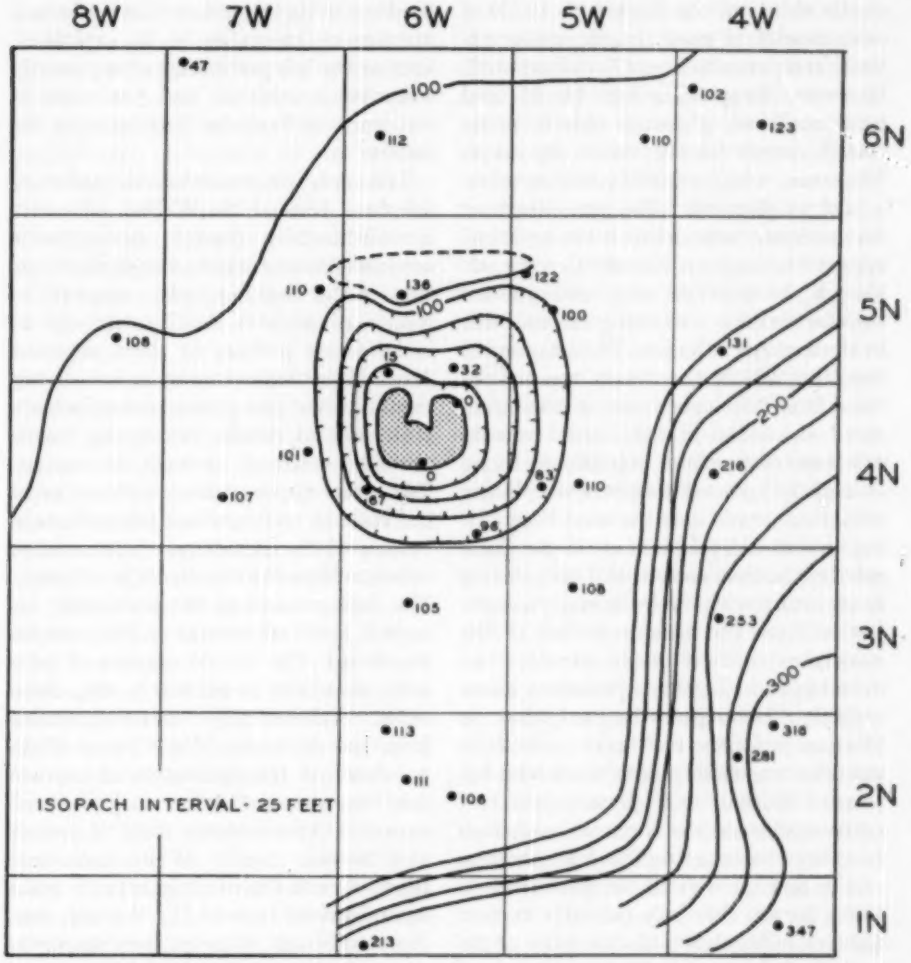


FIG. 6.—Thickness of normal, predominantly red Moccasin Springs deposits around Marine reef. The extent of the zone of reef-induced turbulence is shown by the reefward thinning of the Moccasin Springs deposits and by excess thickness in the mass settling zone north of the reef.

ly that there is considerable range in grain size, composition, and sorting. Random, vertical changes in composition and degree of sorting are the rule. Coarse, cobbly, coral debris beds may be interbedded with fine-grained crinoidal layers. Varying intensities of water agitation in random succession are implied.

Lecompte's contention that the bioclastic debris of the flank beds in these cases consists of small, fragile, uniformly sized components is not substantiated. However, large, detached blocks and coral heads are generally absent, as he stated, except for the storm deposit at Thornton, which contains blocks up to 4 feet in diameter. The percentage of water-worn coarse debris is low and thus appears to support Lecompte's view, although, in general, only well-rounded skeletal remains are readily recognizable in the average dolomite. This may mean that the bulk of the debris was not derived from the higher part of the structure but, instead, that the burial remains are those of the flank populations living in slightly agitated waters. Only during occasional lowering of the wave base during storms did the remains of the flank dwellers become sorted, and only during heavy storms was there a major contribution from the growth surface of the structure. In view of the structure-induced turbulence, this explanation seems unlikely. Lecompte observed that in Silurian times the reef-borers were nonexistent, one of the points of contrast between Silurian and present-day reef communities. As the borers aid materially today in weakening the dead portions of the reef, as well as the anchorage of living forms, this may partially explain the lack of big debris blocks. Some of the reef dwellers may have been swept off and dumped as soon as the forms died, as is the case of the crinoids which have ligament-bound skeletal elements that disarticulate readily. A protective barrier of the lithothamnion-ridge type would seem to have been absent.

Although the flanks were formed under the influence of reef-induced continuous turbulence and although there is bioclastic debris from the reef, both of which show wave resistance of the structure at the time of its formation, there remain a number of anomalies to be explained, such as the low percentage of apparently water-worn material and the ratio of structure- to flank-dwelling forms in the debris.

Lecompte's statement on the nature of the flank beds of the Wabash structure is substantially correct. Recognizable outwash does constitute a negligible fraction of the flanking beds; compact coquinas of skeletal debris make up an insignificant portion of these deposits. Most of the flanks consist of low-clastic, medium-silty, fine-grained dolomite beds, with skeletal debris, principally fragile crinoids, scattered through the matrix. Two other types of flank beds occur at the Wabash reef; both are subordinate in volume to the normal type but are more common than the coquina type of debris. The first consists of the previously described lacy frameworks of *Stromatactis*-like forms. The second consists of dolomitic siltstones to extremely silty dolomites, lithologically indistinguishable from the enclosing Mississinewa shale. As shown in the northernmost exposed cut, they are coextensive with the reef-enclosing Mississinewa shale (Cumings and Shrock, 1928). While individual types of flank deposit may entirely make up individual beds of the Wabash reef, there are places, principally in the northernmost exposure, where at least three types may occur in an individual bed in a transitional vertical succession. Thus

on the northwesternmost flank exposure certain beds grade upward from normal Mississinewa shale into the dominant low-clastic type flank deposit into the lacy-bedded, reef-core type of deposit. Inclined flank beds are, therefore, complex both in composition and in genesis. In the example cited above, the flank bed was initiated under normal quiet-water interreef conditions and gradually came under the influence of the adjacent growing reef, which modified the sedimentation so that finally the site was favorable for colonization by frame-building organisms (fig. 5). Vertical as well as horizontal expansion of the reef governed the formation of this bed. A similar record is shown by one of the frame-builder-bearing flank beds on the southwestern part of the same exposure, where a thin layer of crinoidal debris cuts obliquely across flank beds (fig. 7). As the *Stromatactis*-like forms underneath the crinoidal layer do not penetrate this blanket, it is evident that compact frame-building was confined originally to the lower flank segment, and its upward growth was terminated by the introduction of crinoidal debris. Mass population of the higher segments of the flank by the frame-builders followed, resulting in an upward and outward expansion of the flank formation. Fluctuations in the rate of sediment accumulation are shown by the response and growth pattern of the *Stromatactis*-like forms. Their lacy pattern of growth was produced by alternation of narrow-based, rapid upward expansion with horizontal spread in the form of sediment encrustation. This growth pattern is an index of marginal habitat conditions with reference to sediment tolerance. The columnar restrictions of the growth surface mark the periods of accelerated sediment accumulation which buried the flat-lying colony surfaces. Vertical growth of the remaining prongs of the colony barely kept ahead of sediment accumulation. The crinoidal layer, then, marks a sudden, brief influx of excessive bioclastic debris which temporarily terminated all organic growth. The sediments of the interframe spaces differ but little from those making up the prevalent type of flank bed composed of dolomites low in clastic content and distinguished by the conspicuous absence of *Stromatactis*-like colonies. The differences, therefore, seem to be determined by differences in gross rates of sedimentation. The rate of deposition of the prevalent type of flank bed was at least slightly in excess of the mud tolerance of the *Stromatactis* forms. It was slightly more rapid than that of the type of flank deposits bearing the frame-builders, as the "articulation index" of dismembered crinoids implies (Lowenstam, 1948*a*). The low content of terrigenous clastic material in the flank beds, as compared with that in the coexistent interreef deposits and their flanking tongues, testifies to the more rapid rate of deposition of the former. The latter were laid down under quiet-water conditions and lack both builders and crinoids. It seems obvious that the rates of deposition are primarily a function of the degree of turbulence. The low but persistent content of fine terrigenous clastic material, principally silt but with some clay, suggests slightly, but intermittently, turbulent water for both types of flank beds, with a slightly higher turbulence for those containing *Stromatactis*-like forms. A decrease in terrigenous clastic material, however, is not entirely a function of the intensity of turbulence. Carbonate detritus, slightly larger in grain size than the carrying capacity of the water, supplied by the growing reef, would "dilute" the terrigenous clastic material. Although dolomiti-

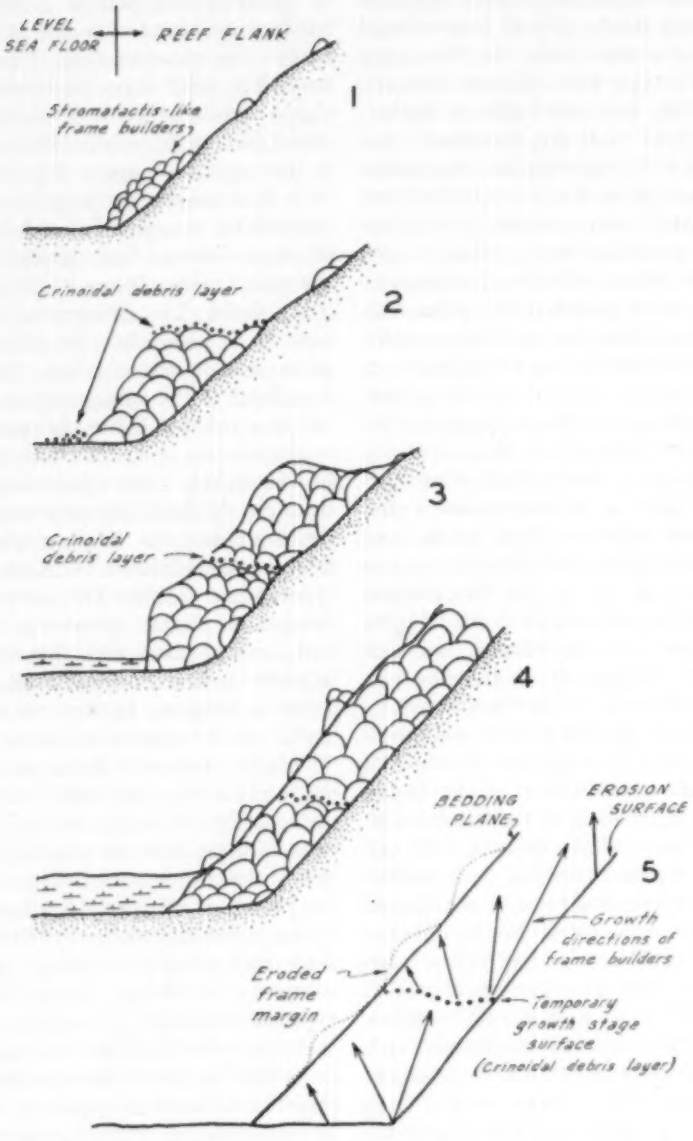


FIG. 7.—Diagrammatic sketch of the growth of reef-flank beds formed by *Stromatactis*-like frame-builders under semiquiet-water conditions; based on data from the Wabash reef, Indiana.

zation has obscured the texture of the carbonate fraction, there is some evidence to substantiate this possibility. The fluctuations in depositional rates, as inferred in the flank bed having *Stromatactis*-like forms, and the recurrence of thin, graded crinoidal coquinas constitute such evidence. A slight increase in turbulence would tend to deflate the reef of finer skeletal debris, leading to a relative reduction in amount of terrigenous clastic material. The crinoidal debris, although quantitatively small, is unquestionably sediment contributed by the reef. The environmental significance of "terrigenous clastic ratio" is deceptive. This is also shown for the Mississinewa shale that borders the structure. That anaerobic conditions prevailed here intermittently is shown by the carbonaceous content, by graptolite remains, and by the organic films mantling the casts and molds of dissolved carbonate skeletons. Weak organic acids, produced by reducing bacteria under the anaerobic to semianaerobic bottom conditions, would tend to dissolve appreciable quantities of calcium carbonate. Thus the present ratio of terrigenous clastic to carbonate is not a reliable index. This ratio has been shifted in favor of the terrigenous clastics by the solution of carbonate.

The four types of reef-flank deposits form a graded series ranging from quiet-water, high-clastic deposits through semiquiet, low-clastic deposits to slightly more agitated *Stromatactis* frame-bearing deposits, with still lower clastic content, to rough-water bioclastic deposits. Genetically, the first is strictly interreef; the *Stromatactis* beds are core type, and the two remaining types range from reef-controlled interreef type to a strictly reef detritus deposit. If reef outwash constitutes a subordinate proportion of the exposed flank segment, it is evident that the reef correlative with it was probably below effective wave base. Prominent contributions of reef debris occurred only during storms which lowered wave base. The fact that the lower portion of the reef core consists largely of lacy *Stromatactis* framework with interstitial low-clastic, crinoid-bearing sediments, such as are found in the *Stromatactis* frame-bearing flank beds, is corroborative evidence. Emergence above effective wave base is indicated by the southward expansion of the highest portion of the reef core over its own flank beds and the correlative development of the reef-core type of flank bed. Details of this will be discussed in connection with wave-controlled reef expansion.

The exposed part of the Wabash reef illustrates a number of critical points. It is evident that the segment of the reef flank is basically a structural element which, unless qualified, does not carry any particular genetic connotation. Like Cumings and Shrock (1928), we have considered the flank beds to be principally reef outwash with minor contributions from reef-builders and reef dwellers living *in situ* and containing tongues of interreef sediments. Although this is commonly the case, the Wabash reef shows that, prior to the wave-resistant phase, the bioclastic contribution of reef origin is subordinate and that the flank may even consist of bedded, core-type rock, especially during the period of upward growth below wave base.

The large Niagaran reefs that border the Ozarks in Illinois, discovered within the last ten years by drilling, show surfaces only slightly modified by erosion, as compared with the more or less deeply truncated, flank-bearing reefs that outcrop farther north (Lowenstam and Dubois, 1946; Lowenstam, 1948*b*, 1949*a*). We are now better able to investigate the

wave-resistant qualities of Niagaran reefs, as it is the reef surfaces of the present day that furnish the criteria for establishment of wave resistance.

Information on the reef surface is most complete for the Marine structure, the gross outlines and topographic relief of which have been delineated by the numerous oil wells that penetrate its surface and by a smaller number of wells lying beyond its border and penetrating the surrounding strata (pl. 5). In all but one well, bioclastic debris was found to form the surface deposits, whereas none was encountered in the wells bordering the reef. The detrital blanket is certainly confined to the reef surface and is a distinct reef element, not a regional lithologic unit. As the detrital mantle is dolomitized only at depth, the composition of the calcareous deposits is satisfactorily ascertained (Lowenstam and Dubois, 1946; Lowenstam, 1948*b*). The thickness of the reef surface deposit varies from 2 to more than 40 feet; but, because the detrital mantle grades imperceptibly downward into dolomite, the thicknesses recorded may actually be only the depth of dolomitization rather than true thicknesses. The few indistinct bedding planes observed in cores are horizontal in the upper part and are inclined up to 20° in the lower part of the generally massive caprock. Apparently, the original topography of the reef surface became gradually buried and to a certain degree leveled off by the debris accumulation.

The mantling debris, wholly of bioclastic origin save for a clear calcitic matrix in certain coarse-grained portions, lacks any terrigenous clastic material. Remains of robust crinoids are the dominant constituent of the coquinas; bryozoan fragments generally make up a subordinate, although locally conspicuous, fraction; the balance is composed of corals, principally tabulates, together with stromatoporoids, brachiopods, and gastropods. The reef-dwelling crinoid genus *Crotalocrinites*, the typical reef coral *P. guelphensis*, and the robust, heavy skeletal nature of the constituents of the coquina, together with the absence of all the indigenous faunal elements of the bordering interreef deposits, show that the debris-forming organisms belonged to the reef community. Yet the percentage of reef-building forms is insignificant; the corals and stromatoporoids make up less than 5 per cent of the whole coquina. Furthermore, reef-core rock has been encountered only in a single well at the southwestern edge of the reef, showing that the debris mantled essentially the entire reef surface. It was for this reason that the writer (1946, 1948*b*) believed this deposit to be a rough-water product formed after cessation of reef-building—in other words, recording conversion from reef to bank development. But is this interpretation valid? If the cause of the conversion was the establishment of a rate of subsidence in excess of the rate of upward growth of the reef-builders, reef-building could have continued, nevertheless, until the reef surface was below effective wave base. Lowering below effective wave base can be excluded, as it would have permitted settling of terrigenous clastic ma-

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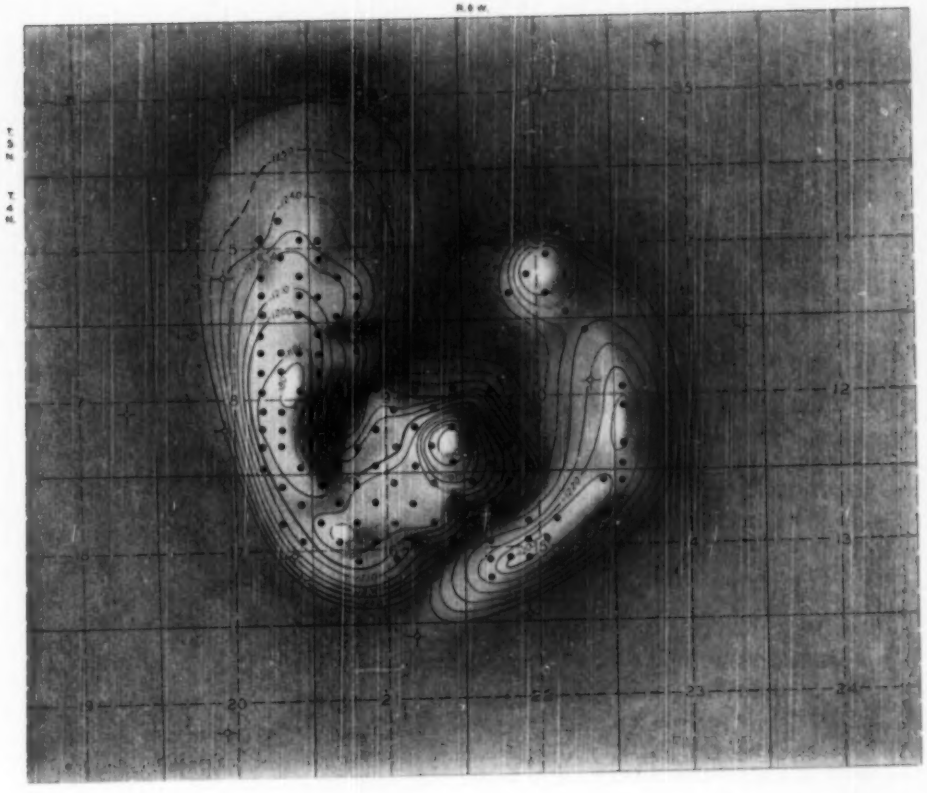
PLATE 5

Outline and topographic relief of the slightly erosion-modified reef surface of the Marine reef, Madison County, Illinois. From Lowenstam (1948*a*, fig. 11), by permission of the American Association of Petroleum Geologists.



FEBRUARY 15, 1947

ILLINOIS STATE GEOLOGICAL SURVEY

DRY HOLE

PRODUCING WELL

CONTOUR ON TOP OF SILURIAN; INTERVAL 10 FEET; DATUM SEA-LEVEL

terial present in suspension in appreciable volume in the Niagaran sea bordering the Ozarks. Subsidence into less agitated water should have resulted in the formation of less, rather than more, detritus.

The apparent reduction in reef-builders might be caused by the establishment of an unfavorable environment due, for example, to lowering of temperature, which would greatly retard reef-building activity. If this were true, we should find a change in the composition of the faunal assemblage. A bank fauna would replace the reef-builders. Inasmuch as the fossil composition and ratio of constituents are the same in the reef flank and the reef surface, we may exclude this possibility and safely conclude that favorable conditions for reef-building must have been maintained.

The only alternative left is that the reef-builders raised the reef to the water surface. It is at this stage in present-day reefs that the volume of wave-dislodged detritus, initially limited and confined to depressions in the reef surface, increases, encroaches on, and gradually takes over the reef surface, finally building it up above water level. At the island-forming stage, reef-building activity and the living reef community as a whole become restricted to the fringes of their former growth surface. It is also important to note that the percentage of reef-builders in the sands covering the reef generally ranges from a common figure of less than 10 per cent to a rare maximum of 50 per cent. The bioclastic mantle of the Marine reef is a perfect fossil analogue of the modern reef surface. Its deficiency in reef-building elements is therefore no anomaly. As in the detrital flanks, the crinoids are the prime contributors to the debris cover. They are the primary source of bioclastic detritus and play the same role that *Halimeda*, mollusks, and Foraminifera do on the present-day reefs. Thus the impression gained that reef-building ceased is still correct, but only as far as the central part of the reef surface is concerned. Evidence for the marginally restricted living fringe with its active reef is furnished by the occurrence of typical core-rock dolomite at the extreme southwestern edge of the reef surface. Although reef-building elements have not been observed in the well cuttings, the fact that the detrital blanket nowhere else was found to be dolomitized is evidence against its being dolomitized bioclastic debris and is consistent with the observed coincidence of dolomite and reef core. It may, nevertheless, seem surprising that there should be only a single record of the actively growing reef fringe. The probability of encountering the narrow growth edge in widely spaced borings is comparatively low, and, more important, only part of the sample records of wells drilled in the critical southern border area were available for examination. On the other hand, closer-spaced samples, available for most of the northern, semilagoon border of the reef showed no trace of reef-core centers at the points penetrated. Patchy development or even absence of an active reef fringe here appears likely in the light of the effect of the prevailing southern winds on reef shape and growth. The reef-core rock encountered 30 feet beneath the detrital mantle in the southwesternmost boring on the fore-reef ridge (discovery well, Eason-Mayer No. 1) contains abundant reef-builders and was previously considered (Lowenstam and Dubois, 1946; Lowenstam, 1948*b*) to be core rock, of an isolated small fore-reef that grew on the detrital reef flank until it was buried by the mantle of bioclastic debris. This rock can now be better interpreted as part of

the continuous rim of the obliquely upward-slanting reef core that projects southward over the flank beds. In other words, this core section was the restricted growth edge of the reef at an early stage. A quite different growth pattern for the Marine reef is implied (fig. 8). Because there is evidence that the reef started in the semiquiet waters of St. Clair time (Lowenstam, 1948*b*), the pattern of development of the Wabash reef is a guide to our understanding of the early growth characteristics of the Marine and other reefs. Growth was initially over the entire reef surface. As the reef approached the water surface, the narrow, steep-walled reef expanded not only vertically but also laterally by peripheral accumulation of reef detritus. The active growth surface, first continuous and centrally located, became patchy, then expanded over the accumulating reef flanks to become, when wave resistance was finally attained, restricted to the reef fringes because of the centrally accumulating skeletal wastes. The detrital reef blanket was thus correlative with the growing reef edge and not younger, as previously assumed.

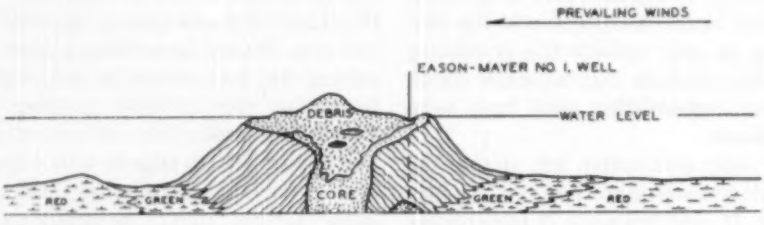


FIG. 8.—Diagram showing reinterpretation of Marine reef at wave-resistant stage, based on the theory of prevailing wind control.

Although the Marine reef had wave-resistant qualities, did it actually emerge above water? From the limited erosion implied by the isopach map of the overlying Wapsipinicon formation (Lowenstam, 1948*b*, fig. 6), the centrally located highest part of the reef complex should have stood about 60 feet above the fringing growth border, strongly suggesting emergence for this part of the reef. If this part was actually above water, calcite cementation was the only available agent for fixation of the loose bioclastic sand, as the oldest land plants known at present occur in the Upper Silurian. There is no evidence of soil formation, unless the pink crinoids, conspicuously concentrated in the highest central reef portion of the blanket, derived their coloration from adsorption of *terra rosa*.

Wave resistance, demonstrated by the Marine reef, is also indicated in the case of the three frontal-reef bastions, McKinley, Bartelso, and Sandoval, where wells penetrating the reef also encountered bioclastic, prevalently crinoidal debris on top of the structures. Because all outcropping reefs, where emergence above wave base is proved by extensive reef-flank deposits, are more or less deeply truncated, the character of their original surfaces is unknown. Wave resistance may have been more commonly attained than can be proved.

The best evidence of wave resistance is the record of a distinct zone of turbulence around the Marine reef during the deposition of the red Moccasin Springs sediments (Lowenstam, 1948*b;* fig. 6).

Similar phenomena have been observed around Bartelso and other less-well-known reefs (Lowenstam, 1949*a*). The zone of turbulence is marked by a halo, principally sandy reef outwash, which replaces the red terrigenous clastic sediments. The clay in the reef-influenced zone is green, unlike the red ferric form found outside the turbulent zone, and is thought to show that the Ozark *terra rosa* became reduced by bacterial action on the bottom, which was richer in organic matter and which was built up by the reef detritus to the euphotic ceiling. Outside the reef area, it remained red because of biologic semideep-sea conditions, lying just below the euphotic ceiling, where a low organic content was coupled with minimum reducing bacterial action. The Moccasin Springs deposits are, in general, thought to delineate a deeper basin lying east of the Ozarks during Niagaran time. Its northern wedge-edge, which lay across half the low-clastic, reef-bearing belt, marked the initial basin slope that later became incorporated into the shelf largely by means of reef-contributed outwash which raised this area above the euphotic ceiling. It can be shown that during the time of Moccasin Springs deposition the radius of the turbulent zone around Marine progressively increased until, at the end of Moccasin Springs deposition, it had reached a diameter measured in miles (fig. 6). Further expansion during the deposition of the Thorn sediments (Lowenstam, 1949*a*) is indicated by coarse reef outwash beyond its previous border. The ultimate radius, less readily traceable in the Thorn deposits, has not as yet been determined. As shown by the isopach sketch map (fig. 6), a narrow, although as yet ill-defined, zone of abnormally thick Moccasin Springs deposits marks the northern fringe of the turbulent zone, a feature also observed north of the Bartelso and a number of other reefs, principally those along the northwestern border of the Vermilion County embayment. These thickened wedges of Moccasin Springs constitute the zone of mass settling on the lee side of reefs, in which fine terrigenous clastics that by-passed the turbulent zone settled together with reef outwash. This tract was, however, not built up above the euphotic ceiling, as was the shallow platform within the turbulent zone. The pattern of expansion of the turbulent zone is exactly what we would expect to find around an expanding wave-resistant reef of the dimensions of Marine, covering about 6 square miles at its final growth stage. This is the most convincing evidence that the Niagaran structures were capable of attaining wave resistance. The complex shape of the Marine reef further disproves Lecompte's claim that the Niagaran structures lacked a form comparable to that of present-day reefs.

The sediments in the low-clastic belt back of the front of the reef archipelago are about double the thickness of those in the reef-free, high-clastic belt to the south. This is an impressive measure of the volume of reef materials contributed to the interreef bottoms and to a narrow strip north of the front of the archipelago. For these reasons the term "interreef" has been redefined by the writer (1949*b*) to denote those deposits which received sediment from the reefs in addition to the regional materials. The surrounding reefs clearly controlled the environment of sedimentation.

In summary, it is clear that Roll and Lecompte err in believing that the Niagaran structures did not attain wave resistance and thus were not true reefs. The reef forms, such as those shown by Marine, the bioclastic debris accumula-

tions similar to those of present-day reefs, and the influence of sedimentation in the environs are convincing evidence of wave resistance. However, the detrital reef flanks pose some as yet unsolved problems and require further study.

PREVAILING WAVE AND CURRENT DIRECTIONS

Wherever recent, wave-resistant reefs grow in the unobstructed path of prevailing winds, the reef forms show the modeling effect of the wind-induced waves and currents. The subject has been dealt with by Umbgrove (1930) and Krempf (1930) and more recently summarized by Yonge (1940).

To secure criteria for recognition of similar effects in ancient reefs, we must study present-day emergent reefs that lie outside the range of island or shoreline interference. The principal effects of prevailing waves and currents on present-day reefs are the growth response of the reef-building organisms and the depositional pattern of reef-derived detritus. Prevailing winds today control the most active growth of the reef-builders, primarily lithothamnioid algae, which form the so-called "lithothamnion ridge." Reef material dislodged and broken off on the windward side becomes effectively sorted by the destructive forces of waves and currents. The coarse material—the coral shingle—is generally piled up on the windward side; the finer material—the reef sand—is carried farther to the lee side. Crescent-shaped reef bodies are the resultant forms in the initial and least complex cases of simple trade-wind control, such as is found in the Cocos-Keeling group in the Indian Ocean (Umbgrove, 1930, p. 49). Reefs growing in areas of strong currents, such as the Thousand Islands in the Sunda Straits, show narrow, elongated forms whose axes parallel the current direction (Umbgrove, 1949, pp. 8, 11).

Application to fossil reefs is obvious. The reef form and the growth trends of the reef core, indicating the direction of most active development of the reef-builders in space, and the time, coarseness, and distribution pattern of reef-derived detritus are indicative of prevailing wave and current directions. In order that winds may effectively control the reef development, two prerequisites have to be met: (1) the reefs must be wave-resistant, and (2) the reefs must be situated in the unobstructed path of the prevailing winds. The latter point is critical, as Umbgrove (1930, 1949) has pointed out in the case of the East Indies reefs, where shielding action by islands, depending on their position with reference to the monsoons, partially cancels or tends to obscure the regional wind relations.

In the Niagaran reef archipelago area, the first of the prerequisites is met. As to the second, the distribution of land relative to the archipelago is known only for its southwestern portion with any reasonable degree of accuracy. Data pertaining to prevailing waves and currents are limited to this area. No land barrier or shifting strand line existed in the central area to separate the north-south expanse of the Niagaran sea between Hudson's Bay and central Tennessee (Lowenstam, 1948*a*, 1949*a*). Two elements within critical distance that may have had shielding effects remain to be considered, namely, the Ozark island and the Cincinnati geanticline. The Ozark island could have effectively shielded reefs proximately located in its northeastern offshore waters, provided that the island was high and the winds were from the southwest. As Foerste (1935) has pointed out, there is no evidence that the Cincinnati geanticline existed as an island in

Niagaran times. The sediments and their fauna that border the structure strongly point toward its existence as a shallow submarine shoal, or swell, whose shifting crest was, most of the time, above effective wave base (Freeman, 1949). Localized, intermittent negative movements lowered portions of it below effective wave base, permitting the entrance of sediments derived from southeastern Appalachian areas into the interior sea. Thus, although the Cincinnati swell could not have formed an effective wind barrier, its effectiveness as a barrier for sediment transport from the east and southeast indicates that it was capable of preventing any current from crossing it. More important, however, is the fact that it must have shortened the distance of fetch of waves from the east or southeast and at least dampened their effect on reefs lying to the northwest. The western and northern approaches to the southwestern portion of the archipelago were apparently entirely unobstructed by land or islands, as was also an open passage between the Ozark island and the Cincinnati swell. These directions were possible avenues of effective wave and current action on the reefs under consideration.

Willis (1910), apparently the first to speculate on Niagaran marine circulation, suggested northerly to northeasterly currents for the area under consideration. This view followed from the then current concepts regarding the Racine-Guelph faunas, which, because they were confined to the northern and northeastern Niagaran, were believed to be northern in origin and derived from the Anglo-Scandinavian basin via Arctic and north Atlantic migration routes. These northern faunas have since been shown to be the reef and rough-water interreef dwellers, and they are, therefore, indicative only of their habitats and not of a faunal province (Lowenstam, 1948*a*, pp. 140–142). The similarity to the Anglo-Scandinavian fauna implies open communications via northern shelf areas suitable for migration. It has no significance in the recognition of current directions.

The form of the Marine reef in Madison County (pl. 5) and the sorting and areal distribution of reef outwash and bypassed terrigenous sediments in the interreef areas surrounding the Marine and several other subsurface reefs led to the view that the Niagaran archipelago portion in Illinois was subjected to southerly winds (Lowenstam, 1949*a*). Additional data and re-examination of previous subsurface information, as well as corroborating evidence from the reefs outcropping in northeastern Illinois and northern Indiana, have materially strengthened the thesis of prevailing southerly winds. The outcrop record has, furthermore, supplied an example of the destructive effect of cyclonic storms hitting the lee side of a reef from the northwest in a manner similar to that observed on recent reefs in the southwest Pacific by Emery, Tracey, and Ladd (1949).

But few reefs showing noticeable effects of prevailing winds are as yet known. The principal effects are the reef form, the direction of core growth, and the depositional pattern of reef-derived detritus. Each of these is considered individually.

REEF FORMS

The buried Marine reef in Madison County, Illinois, and a number of small embryonic reefs in the Lemont area in northeastern Illinois exhibit wind-controlled forms. The Marine reef (Lowenstam and Dubois, 1946; Lowenstam, 1948*b*; pl. 5) is a reef complex about 6 square miles in area, whose surface out-

line in its final growth stage is horseshoe-shaped, with its convex side facing southward. As previously pointed out, extensive reef-flank deposits and the detrital cover blanketing practically all the reef surface show that we are dealing with a partially emerged reef. The two northern horns, enclosing a semilagoon, are highly suggestive of detritus bars that accumulated in the lee of the growing reef front. Because of shallow penetration of the reef surface by wells, this interpretation appears poorly founded. However, recent deepening of a well in the west prong lends support to the interpretation of the detrital nature of this reef portion, as the 122 feet penetrated consist only of reef detritus intertonguing with a thin interreef finger. Thus the southward convex shape of this wave-resistant reef complex and the northward-bending horns, suggesting extensive talus accumulation, are in full agreement with the concept of prevailing southerly winds.

The second example, suggesting current channeling by prevailing winds, is more inferential in nature. Reefs believed to illustrate this effect are located in the very basal Waukesha strata in an intermittently active, shallow flagstone quarry north of Lemont, about 220 miles northeast of the Marine reef. Several of these reefs, exposed at an earlier stage in the quarrying operation, have been described by the writer (1948*a*, pp. 18, 29–33). Six small, narrow-ridged reef bodies are now exposed or are shown to be present by compactional dips of the overlying strata. The reefs show close *en echelon* spacing on the northwest border of a larger, low-lying structure, perhaps a buried Joliet reef. The axes of three of the most clearly delineated reefs trend N. 35° E.; the buried structure runs N. 45° E. The writer has shown elsewhere that the reefs grew in shallow waters just below normal wave base but that they were subjected to intermittent storm action strong enough to overturn and embed, sideways, coral colonies up to 3 feet in diameter. Widely spaced, ripple-marked ridges south of the buried structure and other evidence suggest strong turbulence, which explains why, in Ludwig's quarry, located to the northeast, the basal Waukesha beds are 12 feet thick instead of the regional 25–30 feet. The thinning is due to intermittent by-passing and deflection of the finer clastic fraction. The parallel, linear structures combined with evidence of rough-water conditions strongly suggest a southwest-northeast current which permitted intermittent, short-lived reef growth and controlled the growth form. Recent quantitative studies by Bader on the distribution pattern of the fossil debris and sediment character around one of the minor buried reefs in the flagstone quarry area conclusively show a southwesterly storm control. Because the Lemont area contains the sole record of early Waukesha reefs, it is unlikely that there were enough reefs to produce a shielding action that could have been responsible for the southwesterly wind-induced storm phenomena. The Lemont observations require data from other regions to strengthen the interpretation presented.

Other known reefs do not exhibit forms that suggest molding by southerly winds. The narrow east-west-trending Stony Island structure (fig. 9), possibly a composite of two coalesced reefs, shows too slight an embayment on its northern side to be clearly the result of lee-side concentration of reef detritus. The outlines of the two east-northeast- to northeast-trending frontal reefs, McKinley and Bartelso, are as yet not well defined. The known forms seem to follow pri-

marily the trend of the basin slope from which they rose. Finally, the Sandoval reef, for which a diamond-shaped outline is indicated by the structure of the overlying Devonian strata, is open to opposing interpretations. Depending upon its reef-core form and location, southwesterly or southeasterly wave-modeling could produce the present known reef form. In some cases, such as the Stony Island reef, which was located well inside the archipelago and developed at the time of maximum reef density, shielding by other reefs would reduce the area of fetch and thus could account for reduced wave-channeling of reef sand toward the lee side. However, this cannot be regarded as a factor in the case of the outer reef bastions.

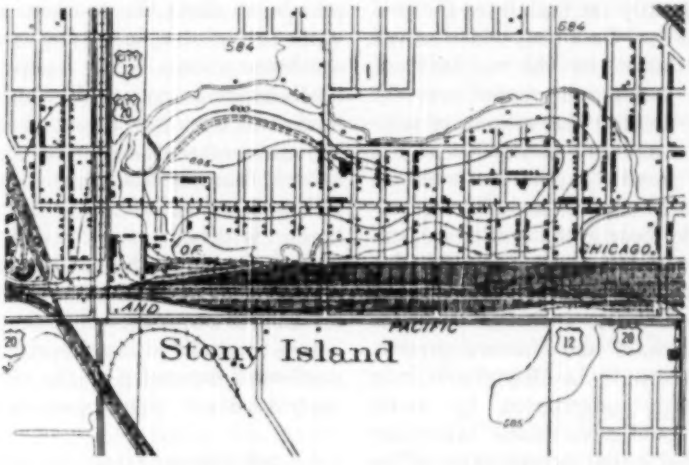


FIG. 9.—Topographic expression of Stony Island reef in Chicago

REEF-CORE EXPANSION

At the Big Four station at Wabash, Indiana, the reef core extends from top to bottom in the northwest-facing railroad cut but is seen to project over flank beds with intertonguing, interreef wedges in the southeast- and southward-lying exposures. Also, as previously noted, the southern outermost flank beds preserved consist of the reef-core type of lattice, reef-builders that genetically represent reef core. A southward reef-core expansion in time and space is thus clearly demonstrated. Furthermore, there is evidence that shows why this directional expansion took place. Unlike the northern, earlier reef core, the higher corelike overgrowth is characterized by a richer, more varied assemblage of reef dwellers, a greater profusion of massive reef-builders, and a negligible content of terrigenous clastic material. Initially, therefore, most of the reef surface was just below wave base, as demonstrated by the comparatively high content of terrigenous material in the interframe spaces of the core, the widely spaced, very narrow frame-builder ribbons, and the paucity of reef-dwelling species. As soon as the reef surface had emerged marginally above wave base, the turbulence reduced the incidence of burial of the reef dwellers by

sediment to a minimum. In turn, it created and stimulated their rapid and more massive growth. A more favorable habitat was created for reef dwellers, as these were sensitive to the excess mud, and materially aided in hastening the upward expansion of the reef. At the same time, the reef came under the influence of the waves induced by the prevailing wind, which stimulated expansion of the actively growing reef front on the windward side. Rapid advance in this direction was greatly facilitated by the preexisting detrital flank beds which served as a substratum for the reef-builders. Moreover, turbulence created over the reef front reduced the volume of sediment that could settle here sufficiently to stimulate growth of the reef-builders. The growth forms of these builders closely resemble those in the border areas of the lower reef-core section deposited before the reef became initially wave-resistant, indicating a similar marginal habitat. Upward and outward growth, although shown by bedding planes to be intermittently interrupted by storm scouring, could nevertheless take place and resulted in the incorporation of the previous unconsolidated detrital flank into the reef frame.

In critically examining the evidence and interpretation just presented, an alternative interpretation has to be considered. The southeasterly portion of the reef alone is accessible to observation. Thus the southward expansion observed higher up in the central reef-core section could actually constitute only the observable part of a radially spreading central reef core. Such an expansive trend would be feasible even under prevailing southerly winds as the result of shielding action by other near-by wave-resistant reefs and is not improbable, considering the great reef density demonstrated by Cumings and Shrock (1928) for this time and place. Also, rapid reef-core expansion of this type has been demonstrated by Hadding (1941, fig. 50) for the mid-Silurian reefs of Gotland, Sweden. Lack of exposure of the northwest portion of the reef core makes final decision impossible. Whether or not the first interpretation is correct hinges, then, on whether or not there is unquestionable evidence of southerly wind control. The reef-core growth of the bedded-flank type is most prominent along the southern margins of the reef klintar, decreasing toward the southwest, none being found on the northeastern or even on the southeastern flank. It is thus only around the projecting outermost southern border that wave-induced frame-builder growth firmly consolidated and actively projected the reef border. Windward exposure and wave effects thus assured, the likelihood of radial spread of the central reef core becomes unlikely at the stage when the reef becomes initially wave-resistant. The southward expansion of the observable central reef core thus appears real.

DEPOSITIONAL PATTERN OF REEF-DERIVED DETRITUS

The finest reef outwash, the reef milk, would be carried by currents created by prevailing winds and would settle in sizable quantities past the zone of reef-induced turbulence. The position of the zone of maximum sedimentation would thus be an additional index of the prevailing wind direction. This zone at the Marine reef occurs north of the reef, but only the trend of its eastern limb has so far been determined (fig. 6). A western extension is indicated, however, by a thickness of the Moccasin Springs formation slightly greater than the regional average. At the Bartelso reef, only the northeastern portion of the zone of mass

settling has so far been located (Lowenstam, 1949*a*, fig. 5). Similar lee accumulations of several other subsurface reefs are known, and in each case they lie near the northern borders, thus corroborating other evidence of prevailing winds. The extent and precise configuration of these zones with reference to the reef diameters is only partially known. The meager data at hand indicate their projection beyond the width of the controlling reefs, as at Marine, which suggests a wind direction variable from the southeast to the southwest.

Observation on present-day reefs has shown that storms approaching them from the lee cause major destruction on the steep and even overhanging reef edge. Large portions of the reef edge become dislodged and are carried down the slopes (Emery, Tracey, and Ladd, 1949). The record of a similar event during Niagaran times is found in the southern quarry at Thornton, where a breccia bed can be traced from the erosion-bisected flank of the main reef more than halfway around the south and southeast face, where it disappears below the quarry floor. This bed is made up of a variety of unsorted, variously sized and shaped pebbles and blocks up to more than 4 feet in diameter, consisting of reef rock, detrital flank rock, and shale pebbles in an argillaceous, fossiliferous, dolomitic matrix, grading locally into a dolomitic siltstone. The number and size of the fragments generally decrease toward the top of the unit. The breccia bed marginally overrides the edge of the lower southwestward-lying portion of a steep-walled fore-reef in the southern quarry face. The bed tongues out within 10 feet east of the structure; the interspace is filled with a dolomitic mudstone. A more centrally located fore-reef, a part of which is left in the same quarry, has the conglomerate bed in contact with its western side, where it contains large blocks of reef core, while its eastern to northeastern border is again conglomerate-free. The eastern and southeastern conglomerate-free edges of the two fore-reefs establish the direction from which the conglomerates came; the reef materials within it establish its source. A storm approaching the main reef at Thornton from the northwest, the lee-side area, was responsible for the most destructive event that befell the reef. The argillaceous interreef beds of the lower flanks were evidently involved, producing a mud flow. Reef rock blocks torn loose, reef and reworked interreef fossils, and semiconsolidated interreef sediments were incorporated as the flow moved downward, surrounded, and partially moved up lower fore-reef portions, parted around their lee sides, and merged again a short distance past them. Specific details will be found in the forthcoming report on the Silurian stratigraphy of northeastern Illinois by Willman, Workman, and the writer.

THE REEF FOUNDATION

The observation that many present-day reefs rise from hard bottoms led to the conviction that a consolidated platform is essential for a reef foundation. It is now known that reefs may start on unconsolidated bottoms, although the number of such cases appears relatively small. The patch reefs of Batavia Bay are examples of mud-rooting reefs, and similar reefs have been reported from other parts of the East Indies (Umbgrove, 1949). Reef-building organisms, nevertheless, require a solid surface for attachment, and, in the absence of a suitable platform, pumice, pebbles, and hard skeletal remains have served as objects for attachment of the pioneer populations.

Cumings (1932) was the first to con-

sider the nature of the Niagaran reef foundations. In the absence of exposure of the actual reef base, the nature of the substratum was deduced from the character of the surrounding sediments and the structures attributed to settling of certain reefs into the substratum.

The Niagaran reefs started their growth and spread throughout the Great Lakes area well after the beginning of Niagaran sedimentation. Nowhere is there evidence of a hard substrate within this region. The potential reef foundations range from argillaceous carbonate muds to sands of bioclastic debris. Only the latter occurs in the clastic-free belt.

The nature of the substrate can be demonstrated where reef bases have been observed in outcrops. Reefs resting on argillaceous dolomites, that is, muddy-bottom deposits, are exposed at Wauwatosa, Wisconsin, in the Schoonmaker quarry (Shrock, 1939), and in the Federal quarry west of Brookfield, Illinois, where a small reef rests on highly argillaceous, cherty dolomite. The pioneer colonies underneath the quarried-out reef core at Elmhurst are surrounded by argillaceous, cherty dolomite. Reefs resting on sands of bioclastic debris are exposed in Illinois in Ludwig's quarry north of Lemont, in the west face of the Federal quarry west of Brookfield, in the southern flank beds of the Thornton main reef, and in the north-facing bluff east of Cordova. A sediment of intermediate composition is found underlying a small reef in the unnamed flagstone quarry north of Lemont (Lowenstam, 1948*a*).

The settling structures of Niagaran reefs, first noted by Cumings and Shrock (1928) in northern Indiana, are equally well shown in northeastern Illinois by the Elmhurst reef and a small fore-reef at Thornton (Lowenstam, 1949*a*). That this settling took place during the reef growth, and not afterward by compaction, is well shown by the Elmhurst reef, where, as a result of the settling, a shallow depression formed in the area of contact of the reef flank with the interreef bottoms. Stagnant anaerobic conditions prevailed intermittently in the depression and, as a result, selectively preserved organic detritus, including graptolites, the remains of which are not found elsewhere around the reef. Reef settling has been demonstrated only in the low-clastic belt, where the substrate was argillaceous, carbonate mud. The bottoms, therefore, not only were unconsolidated but, as a result of loading, were plastically deformed as well.

The bottom conditions were evidently not a hindrance to luxurious reef growth, as the size of the individual reefs, as well as their areal spreading, indicate. Solid objects must then have been abundant to enable the pioneer populations of so many reefs on the bottom muds and sands to secure a footing. Skeletal remains and, following the formation of a wave-resistant reef, dislodged reef pebbles and blocks constitute the only supply of potential attachment objects. The only coarse, reef-derived debris observed to date—the storm conglomerate at Thornton—apparently was not the substratum of any fore-reef. The exposed reef bases have not yet shown that skeletal debris actually formed the foundation materials of these reefs. The contact with the underlying sediments shows only a gradual transgression of the reef base from the center outward over progressively younger beds without any evidence of bottom-rooting reef-builders. The randomly exposed sections of reef cores apparently do not transect the initial growth center. The pioneer population of the reef at Elmhurst, consisting of large domal colonies of *Syringopora*, is

not well enough exposed to show its attachment to the substrate. Although there is little question as to the unconsolidated nature of the bottoms from which the reefs rose, there is as yet no observed record of the objects of attachment for the pioneer population.

Environmental controls other than the availability of attachment surface must have determined the sites of initial reef development. The *en echelon* arrangement and density of the small reefs in the flagstone quarry north of Lemont, as pointed out above, suggest current control in the selection of their building site. The *Syringopora* colonies of the pioneer population of the Elmhurst reef show a similar northeast-southwest alignment (N. 27° E.) of the elongated axis of the individual colonies and of their spacing. An increase in the bryozoan ratio in the interreef assemblage surrounding the colonies, which is thought to mark a transition from perfectly quiet- to semi-quiet-water conditions, suggests a slightly stronger circulation over the future reef area. A slight reduction in volume of the terrigenous clastics that settled out may have been the determining factor in establishing the *Syringopora* colonies that initiated the Elmhurst reef. Although increase in circulation and the resultant reduction in volume of settling clastic materials may have been very small, it might have been significant in terms of the threshold of tolerance of the *Syringopora* forms. Evidence of marginal conditions for the reef-builders in the initial quiet-water stage of the Wabash reef is shown by the expanding and contracting frame-builder surfaces. The abrupt termination of growth of a fore-reef at the southern periphery of the Thornton reef was apparently due to a slight increase in volume of settling mud. The reef-covering bed contains more terrigenous clastic material than do those bordering the structure.

The sudden change in volume of the clastic content common in the reef-bearing sections cannot be related to changes in regional supply of sediment. Changes in local reef geography, primarily of wave-resistant reefs, that caused shifts in the zones of turbulence and modified the interference pattern of closely spaced reefs, which resulted in current diversions, is the more probable cause of such abrupt changes in the character of the sediment.

There are other causes of abortive, short-lived reefs, such as burial by reef outwash from adjacent reefs. The subsidiary small reefs enclosed in bioclastic debris of the southern flank of the Thornton reef and a small reef mass in detritus bed No. 7 at Ludwig's quarry (Lowenstam, 1948*a*) are examples. Although the zone of reef-induced turbulence which prevents settling of fine terrigenous clastics initially favors the growth of secondary reef centers, an increase in volume of bioclastic debris from the primary reef surface ultimately leads to burial of the secondary structure. The greater the distance from the controlling center, the greater the probability that the secondary reefs would attain wave resistance, whereas the more closely located building sites had a lower chance of survival because of encroachment by excessive reef debris.

From the foregoing it appears that the probability of reef formation becomes progressively smaller with increasing muddiness of the sea and also with greater bottom depth. There is some evidence that the reef density was lower in the area of the Moccasin Springs deposition than it was around the head of the Vermilion County embayment and in the outcrop areas in northeastern Illinois and

northern Indiana, where the Moccasin Springs is absent. Yet there seems to be an apparent contradiction in the fact that the reefs with largest areal extent and maximum aggregate thickness found to date lie within the area of the muddy Moccasin Springs deposition. What, then, are the factors controlling the initial location and the development which culminated in wave resistance?

The few wells that penetrate the Marine, McKinley, Bartelso, and Sandoval reefs supply some information, particularly on the problem of the initial site. In every case the reef portions penetrated are found to rest on sections of Moccasin Springs of varying thickness, amounting to 35 and 60 feet at Marine, 230 feet at McKinley, 116 feet at Bartelso, and 35 feet at Sandoval. Thin wedges of the Moccasin Springs beds occur higher up in the McKinley reef; much interfingering with reef rock marks the basal 245 feet of the reef section at Sandoval. These Moccasin Springs deposits show little or no resemblance to the normal formation, save in the McKinley section, where a single 10-foot-thick tongue, some 17 feet above the base of the formation, consists of the usual red and purplish siltstones and silty limestones. These atypical Moccasin Springs sediments are variable, and, as a rule, the terrigenous clastic content is low. They are dominantly or uniformly greenish calcareous siltstones and shales. The normal red and purple colors are subordinate or commonly wanting. Coarse semicoquinas of robust crinoidal skeletal material, sponge remains, and chert, all foreign to the normal Moccasin Springs, may be present.

Elsewhere (Lowenstam, 1948*b*) it has been pointed out that the atypical Moccasin Springs sediments around and underneath the Marine reef are the product of reef-created turbulence and contain reef-derived outwash. The occurrence of these, and only these, reef-modified sediments underneath portions of the reef complex was considered evidence that the reef must have started at a point elsewhere in the complex during the preceding St. Clair time. The Marine reef, which therefore started its growth during the relatively clear-water phase of St. Clair deposition, must have attained a vertical height sufficiently great that it reached effective wave base at least by the beginning of the muddy phase in the succeeding Moccasin Springs time. Bypassing of the fine terrigenous clastics thus became assured, and the turbulence created over the reef and in its immediate vicinity modified the sediments within the influence of the reef. These deposits were subsequently overridden by the expanding reef complex (Lowenstam, 1948*b*).

The occurrence of similar modified Moccasin Springs deposits underneath the McKinley, Bartelso, and Sandoval reefs justifies the conclusion that these three reefs also started their growth during the preceding St. Clair time. The abnormal character of the St. Clair in the Sandoval well lends additional support to this view. As at Marine, the wells through the other three reefs show that they were located somewhat off the center of the initial growth. The variations in thickness of the subreef sections penetrated would suggest that the lines of penetration were not all equally eccentric. These variations in thickness cannot be readily translated into differences in the time required for the reef to expand to the well locations, as the rates of accumulation are a function of the amount of reef outwash reaching these points. The shifting proportions of recognizable reef detritus imply considerable fluctuation in the size of the sphere influenced

by the reefs, particularly in the cases of McKinley and Bartelso reefs. A most striking example is the occurrence of a normal Moccasin Springs tongue at the McKinley reef. The distance from the margin of the growing reef, as well as changes in configuration of the reef, are responsible for shifts in direction and amount of outwash carried to a given place. These shifts are reflected by fluctuations in composition and modification of the normal Moccasin Springs deposits. These changes may also reflect expansion and constriction of the reef surface prior to emergence into the surface waters, after which rapid expansion could and actually did take place.

The objection may be raised that it is too much of a coincidence to find Moccasin Springs deposits (even though modified) underneath the sections of every one of the wells that penetrated reefs if the reef growth started in St. Clair time. To dispel the objection, it may be pointed out that, according to the law of probability, the likelihood of penetrating the initial growth center by one or two wells would appear slim. Even in the Elmhurst reef, the single outcrop example of an observed pioneer population, the initial growth center is unknown.

Corroborating evidence from the outcrop area may, however, be more convincing. The flank beds of the Wabash reef in northern Indiana have been shown to contain terrigenous clastics and reef outwash. It is conceivable that the purer reef detritus-bearing portions of the atypical Moccasin Springs sections underneath the unquestionable reef deposits at McKinley, Bartelso, and Sandoval are actually similar reef-flank portions that accumulated prior to emergence of the reef surfaces into the actively agitated surface waters and that the more argillaceous beds represent interreef overlaps similar to those found at the Wabash reef. Because of lack of cored sections, which should show inclined bedding, the initially presented conservative interpretation of the atypical Moccasin Springs deposits is preferred. In either case the St. Clair age of the initial reef seems most reasonable, indicating only that reefs initiated during the early Niagaran clear-water phase (St. Clair time) were capable of maintaining themselves during the succeeding extremely muddy Dixon phase of Moccasin Springs time. Following the gradual restriction of Moccasin Springs sedimentation, i.e., after the raising of the sea floor from levels below the euphotic ceiling into the lighted zone, and incorporation of these bottoms into the reef-bearing shelf by means of accelerated sediment accumulation through reef-derived outwash, reefs could have been established between the pre-existing ones. Evidence in this as yet too sparingly drilled area is wanting but, if verified, would show several reefs in the upper Thorn wedge sediments.

In all cases, then, there appears to be a definite mud tolerance indicated for the reef-builders. Insoluble residue studies properly interpreted in terms of rates of sedimentation may yield a quantitative approach for this purpose. The basin-bordering reefs also indicate a depth ceiling for initial reef development apparently determined by the depth of the euphotic ceiling.

TOPOGRAPHIC RELATIONS OF THE REEFS TO THEIR SURROUNDING BOTTOMS

Organic structures, if rigidly constructed by frame-building and sediment-building organisms, can, unlike unconsolidated bank accumulations, stand up above their surrounding bottoms without the support of marginally ac-

cumulating sediments and can maintain slopes steeper than the angle of repose of loose sediments. Such structures, true reefs, may rise to a considerable height above their surroundings if their net growth rate, determined by the building rate of the organisms in excess of deflationary processes, exceeds the rate of sediment accumulation in their environs. This relationship holds true only in cases of reefs rising from deep water toward the surface on a stable platform or on a substrate that sinks at rates slower than the increment of growth on the reef. Once the intertidal zone is reached, reef growth shifts entirely from the vertical to the horizontal unless the growth potential is surpassed by the sinking of the substratum.

The rigidity of the Niagaran structures is assured by the frame-building and sediment-building abilities of the reef-builders and is most convincingly demonstrated by the wave-resistant qualities of the reefs. Although the Niagaran reefs were potentially able to rise above their substrata, did they actually do so, and, if they did, what vertical dimensions did they attain?

Shrock (1939, pp. 555, 561) believed that the contact of the Silurian reefs of Wisconsin with surrounding beds indicated that these structures "stood a few tens of feet above the surrounding bottoms." This is the only reference to the subject in the literature, and, although vague as to actual height, it gives an order of magnitude that points toward comparatively low-lying structures. Yet the reef heights indicated are considerably greater than the few meters given by Hadding (1941, p. 55) for similar Silurian reefs of Gotland, Sweden. Of the various contacts cited by Shrock to establish the topographic relief of the Wisconsin reefs, only those of flank-bearing reefs give any assurance that these reefs actually rose above the interreef bottoms during their growth. The talus-like, radially dipping flank beds, wedging out marginally, require some topographic relief for their formation. Yet only a few feet of relief over prolonged periods of time would be sufficient to produce low, impressive talus wedges. Moreover, secondary accentuation of flank-bed dips by compaction would simulate the deposits of reefs of greater height. Conversely, some reefs may have stood above their surrounding talus platforms or "reef foot," as Cumings and Shrock (1928) have suggested for some of the reefs in northern Indiana. To determine accurately the topographic relief of the Niagaran reefs during their growth would require, first of all, knowledge of the time equivalence of specific levels of reef growth to their surrounding interreef strata. In the case of wave-resistant reefs with a lower-lying, fringing, talus platform, there may be discontinuity between a given reef surface and its contemporaneous accumulating reef-flank horizon, the connecting surface element constituting a diastem. Assuming that such data had been obtained, we would have to compute, next, the compaction factor for the interreef deposits, in order to compensate for the present exaggerated differential in levels. A slight correction would also be required for the reef height, as observed brecciation and solution reduction points toward slight volume shrinkage of the reef mass.

Actually, such time-level relations have not as yet been determined, partly because of lack of criteria to link the faint growth lines observable in a few cases of dolomitized reef cores with their equivalent flank segments and interreef beds and partly because of the inadequacy of exposures and the deep truncation of the reefs. Inadequacy of expo-

sures has been the most serious handicap in any attempt to approach the problem, as either the outcropping reef-flank segments are younger than their associated core, as in the case of the Thornton reef, where their top portions have been removed by erosion, or their time equivalence is still not certain, in cases such as the Bluffton reef, where they are seen to merge with the core rock. In the latter case the apparent gradation may be due only to obliteration of the contact by dolomitization. These outcrop difficulties become most apparent where we find time-equivalent key horizons linking the reef flank and interreef, as at the Thornton reef, where the storm deposit noted would have made determination of topographic relations possible, had it not been for the erosional obliteration past the lower extremity of the reef flank. It is for these reasons that reef heights are seldom discussed and that Shrock was not able to present more accurate figures.

At the present moment, there is only one method of evaluating the approximate relative topographic relief of fossil reefs. The relative height of reefs is deduced from the nature of the interreef sediments, that is, whether they were deposited above or below wave base and, in the case of the latter, above or below the euphotic zone. Again, the exact depths of effective wave base cannot be determined, as it would require knowledge of the area of fetch, which, in turn, would require more information on actual spacing of wave-resistant reefs than is now available. From what we know about relative reef density, shielding effects must have increased from the archipelago border inward and possibly varied within, depending upon the relative proximity of reef clusters, their size, and the variations of fetch. Certainly, the wave base was considerably higher than the general figures of between 125 and 150 feet, which probably apply to open, unobstructed waters. If areas inside the outer "north reef barrier" at Bermuda are typical, a depth of 30–50 feet for wave base may have been of the right order of magnitude, as the writer dredged clay-sized carbonate muds at these depths in the Bermuda inshore waters. In the open North Sea, mud deposition starts 4 km. from the coast at a depth of 25 meters, or about 75 feet. Sverdrup, Fleming, and Johnson (1946) give 80 meters, or about 240 feet, as the ceiling of the euphotic zone. Actually, as they pointed out, the depth of this ceiling is a function of the relative clarity of the waters and of the latitude. In the southwestern frontal area of the archipelago, within the area of red-mud deposition of the Moccasin Springs formation (Lowenstam, 1949*a*), considerable muddiness of the waters is indicated by both the fine terrigenous clastics and the character of the fossils (Lowenstam, 1949*b*, p. 115). To correct for the latitude, an average of H. Schmidt's figures (1935, p. 10) of 180 meters for the Mediterranean and 35 meters for the Baltic Sea, that is, slightly over 300 feet, is chosen. Because of the inferred muddiness of the waters, the euphotic ceiling was probably nearer 200 feet. Although inferential and very crude, the values of about 40 feet for effective wave base and about 200 feet for the euphotic ceiling may be adequate for our purposes.

A gradual deepening of the waters from western Iowa and northwestern Illinois southeastward toward the archipelago front in western Illinois is indicated by the character of the interreef deposits. Three zones of differing mean depth may be recognized. The northern one coincides with the "clastic free belt" that extends from western Iowa through

northwestern Illinois into northern Wisconsin. Its southern border zone extends into western Michigan and Ontario. A central zone coincides with that part of the low-clastic belt in Illinois and Indiana and southern Michigan which lies north of the northern wedge-edge of the Moccasin Springs formation. A southern zone coincides with the Moccasin Springs distribution area of the low-clastic belt in southern Illinois and adjacent southwestern Indiana.

The sediments of the clastic-free belt, despite large-scale dolomitization, contain recognizable bioclastic debris, glauconite, and particularly oölites. We know, therefore, that the mean depth of this belt was above effective wave base (Lowenstam, 1949*a*, *b*). Reef heights were therefore generally much less than 30 feet. This view is corroborated by the shallow-bodied, wider-than-high, lenticular reef bodies observed in the northward-facing bluff east of Cordova, at Albany, in northwestern Illinois, and the reef lenticles exposed in several quarries on the Northern Peninsula of Michigan. The partially exposed reefs at Grimm and Quarry (Shrock, 1939) in northern Wisconsin are similar, as is the marginal reef portion at Wiarton, Ontario. The shallowness of the waters prevented the upward growth of these reefs and led to lateral spreading. It is also clear why these reefs lenticles lack discrete flank beds. Giant cross-bedding shows that detritus swept off the reefs was constantly reworked and redeposited on the surrounding bottoms. These reefs, like the Gotland reefs described by Hadding (1941), stood only a few feet, definitely less than 10 feet, above their surrounding bottoms. The intermittent deposition of talus-bearing reefs with discrete talus fans, as found at Port Byron and Cordova in northwestern Illinois, indicates temporary sinking of the sea floor to depths near effective wave base. Average reef heights for the clastic-free belt were therefore of the order of 10 feet, as suggested by Shrock.

The sediments in the second zone indicate a bottom depth a little below effective wave base, with some local upbuilding by reef sand to slightly above wave base, decreasing proportionately from the northwestern to the southwestern part of Illinois. The reefs, mainly wave-resistant, are characterized by prominent talus fans, well illustrated in the eastern half of the Thornton reef. The average heights for the reefs of this central zone were probably of the order of slightly over 40 feet. Total thicknesses of reef sections, penetrated here in subsurface in Piatt, Logan, Livingston, and McLean counties, range up to nearly 200 feet, and, although there is no assurance that the single well penetrations cut the reefs centrally, these figures are closer to actual thicknesses than any available for the partially exposed, truncated, outcropping ones, whose maximum thickness of about 60 feet is seen in the Elmhurst reef. Low-lying flankless reefs are known, however, in the outcrop area of northeastern Illinois. Their flankless nature stems from two different causes. They are either abortive reef growth centers formed on reef sand above wave base on reef flanks or near reefs that were buried early in their growth by shifting bioclastic debris from the near-by wave-resistant reefs, as at Thornton, at Sag Bridge, in the Federal quarry near Brookfield, and in the bluff exposure in Ludwig's quarry north of Lemont; or they are similarly shaped reefs which started to grow from the muddy bottoms below wave base but had their growth terminated before reaching wave base through excessive mud settling, as shown

in a fore-reef in the southern wall of the Thornton quarry. Only the top 20 feet of this reef can be seen in exposure. Its height at any given time of its development was probably of the order of less than 10 feet. This is shown by the height of the southeastern portion, that either continued to grow or stood above the debris flow cover of the storm deposit, previously mentioned, that terminated growth of its southwestern portion. The lack of flank beds in the former type formed above wave base is due to redistribution of the reef-derived detritus; in the latter type, detritus is lacking because all reef growth occurred below effective wave base.

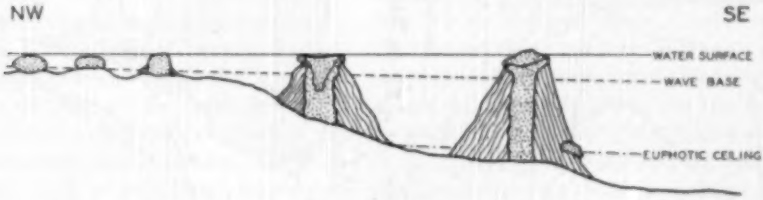


FIG. 10.—Schematic cross section across the reef archipelago portion in Illinois, depicting types and heights of reefs in relation to water depth during Moccasin Springs deposition.

The reefs of their border tract—that is, Marine, McKinley, Bartelso, and Sandoval—and the reef clusters bordering the Vermilion County embayment on the Illinois and Indiana sides are surrounded by strata bordering and interfingering past their zones of turbulence with the red muds of the Moccasin Springs formation (Lowenstam, 1948*b*, 1949*a*). These reefs went through a phase of growth when the bottoms were marginal to the euphotic zone. While the bottoms were built up to slightly below wave base during deposition of the Thorn wedge, the Moccasin Springs interfingering turbulent zone indicates reef heights of the order of about 200 feet, becoming, in the following depositional phase of the Thorn deposits, nearer half their former height. Because of the open waters to the south and to the west in the case of the frontal reefs of southern Indiana, effective wave base must have been considerably lower than the estimated 30 feet within the shallow waters of the reef-dotted interior of the archipelago to the north. If this initial growth of the southwesternmost reef took place in the closing phase of the shallow semirough-water St. Clair deposition, as is strongly implied in the discussion on the reef substrates, a size pattern of initial low-reef form, followed by a tall-reef development and final reduction in height, is indicated for these reefs. A slightly higher maximum height for their intermittent growth maximum is indicated by the greater Moccasin Springs accumulations here, as compared with the reefs southwest and southeast of the Vermilion County embayment. The reefs of this outer deepest zone show total growth aggregates ranging from 600 to almost 1,000 feet (Lowenstam, 1949*a*).

In summary, the average reef height apparently increases from north to south inasmuch as the substratum in the northern areas is generally above wave base whereas that in the south is in general below wave base and in some places below the ceiling of the euphotic zone (fig. 10). Figures as to the estimated reef height indicate primarily the order of

magnitude of the trend in size increase to the southwest, and to the southeast in Indiana. The increase is from an average of about 10 feet in the northwest zone through about a 30–100-foot maximum in the central zone to intermittently up to about 200 feet, but probably averaging 100 feet, at the archipelago front. Total aggregate thicknesses for the reefs, although primarily an index of the rate of sinking of the substrate, is also a relative measure of size relations, increasing along the same trend from less than 10 to slightly less than 1,000 feet.

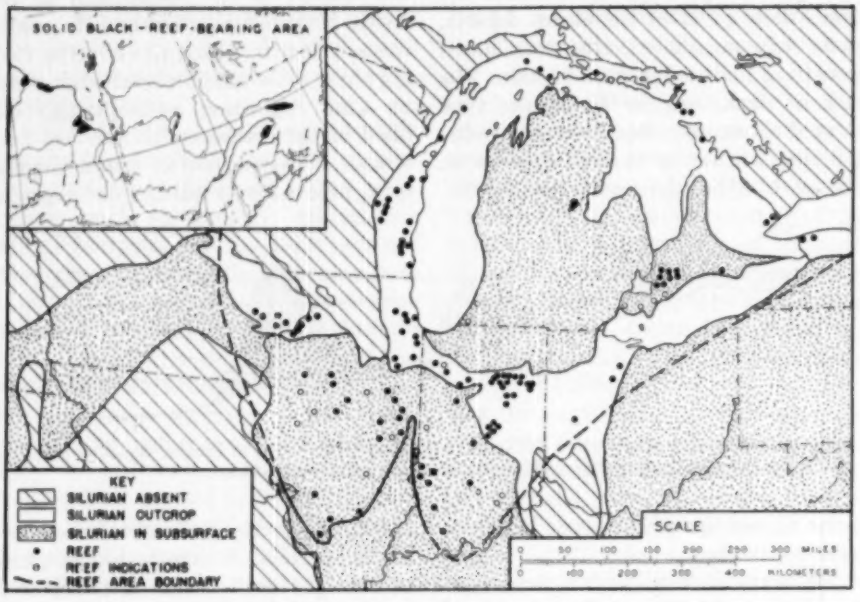


FIG. 11.—Distribution of Silurian strata and location of known reefs

DISTRIBUTION AND LIMITS OF THE REEF ARCHIPELAGO

The distribution of the reef archipelago, as shown in figure 11, is only approximate, inasmuch as the original borders are unknown. Furthermore, it includes both Albian and Niagaran reefs[2] and is therefore a composite. Only on its southwestern border is it accurately defined, and even here additional subsurface information will undoubtedly modify some details. Elsewhere, the original borders have been either removed by post-Niagaran erosion or are as yet unknown, owing to insufficient subsurface data or to lack of study. Although its western, north-

[2] The reference of the Rockhouse of western Tennessee to the Upper Silurian (Cooper, 1942) and the suggested Upper Silurian age of the transition beds between the Bainbridge and the Bailey in southeastern Missouri and southwestern Illinois (Lowenstam, 1949a) point toward continued marine Upper Silurian inundation not only of the Appalachian geosyncline, long recognized, but also of the interior of the continent. If this is correct, the question arises as to whether the development of the reef archipelago terminated at the end of the Niagaran or lasted through the succeeding Cayugan interval. The reef-type fauna of the Cobleskill of New York favors the latter view. Within the archipelago portion under consideration, the Niagaran and any younger Silurian strata that might possibly be present north of the archipelago front underwent uplift

western, and northeastern margins have been mainly obliterated by erosion, we may yet determine its northwestern extension by study of the now buried erosional remnants between the Iowa outcrops and the Silurian of the Forest City Basin and by integration of the subsurface records from North Dakota with data on the Manitoba outcrops, and its northern extension by a detailed study of the exposures in the Hudson's and James Bay areas. Subsurface studies are as yet lacking to determine whether or not the southeastern border can be established beyond the erosional gap of the Cincinnati arch.

Uncertainties of some long-range correlations, particularly of the pre-Guelph formations between beds in the low-clastic belt and those of the clastic-free belt; the doubtful stratigraphic position of some reefs; and the uncertain dating of subsurface reef bases hinder the accurate reconstruction of the pattern of spread of the archipelago in space and time. Nevertheless, the broad outlines of the pattern are discernible. The initial phase begins in early Albian (Manitoulin) time, and the earliest reefs, although small and apparently few in number, are located on Drummond and Manitoulin islands. They occupied a central position with reference to the archipelago, as now known.

Reefs correlative with the early Albian structures have not been found south of the Drummond-Manitoulin line either in the extensive outcrop areas or in subsurface, but they may have been present to the north or northeast. The next younger reefs so far known occur in the Burnt Bluff formation. They consist of the numerous and as yet questionable "bioherms" of Shrock (1940) that occur in the lower Byron division of the Burnt Bluff in Wisconsin, north of Milwaukee. They may, in part, be shallow-water storm concentrations of skeletal debris. One definite small reef is reported from the upper Hendricks division of the Burnt Bluff formation in the Hendricks quarry on the Northern Peninsula of Michigan.[3] If the Byron bioherms prove to be reefs rather than detrital mounds, a southward migration of the reef-building activity would be established. In contrast to the meagerness of actual reefs, there was widespread luxurious coral and stromatoporoid carpeting of the sea floor in late Albian time. These are the coral beds that occur in the "Joliet" formation of Savage (1926) in Iowa and northwestern Illinois (now known to be upper Kankakee [Lowenstam, 1949*a*]), in the basal Kankakee of northeastern Illinois, in the Byron dolomite of Washington Island in Wisconsin, and in the Hendricks division of the Burnt Bluff in the Northern Peninsula of Michigan. These coral beds always contain potential reef-building forms, such as stromatoporoids, *Favosites*, and *Halysites*. The associated gastropod and brachipod fauna contain elements of the type later found among the characteristic reef dwellers, such as a species of *Trimerella*. Certain unknown environmental stimuli needed to induce rapid growth and to produce fusion of the potential reef-builders into solid reef bodies were lacking.

and erosion prior to the Dutch Creek–Geneva inundation. As neither Upper Silurian reef nor interreef fauna have as yet been recognized in the frontal-reef area, where we would expect preservation of erosional remnants if they were present, the possibility of Upper Silurian reef development remains to be determined.

[3] This reef was observed by G. M. Ehlers in the Hendricks quarry in the upper Fiborn strata but has long since been removed by quarry operation. According to Ehlers' personal statement, the reef core exposed was about 20 feet long and from 3 to 4 feet wide and consisted of reef-building corals. It was surrounded by a few ill-defined flank beds.

Intense reef-building started only in the following Niagaran epoch, when the large, wave-resistant reefs came into existence. The reefs generally increased in size, density, and areal distribution with time. By Racine-Guelph time, the density and areal extent of the reefs reached a maximum, and an archipelago that extended from the border of Ozark island northeastward, at least as far as southern Ontario and probably as far as Hudson's Bay, was formed. Outliers may have extended as far south as western Tennessee (Wilson, 1949).

There is insufficient information to determine precisely the initial centers of Niagaran reef growth and the pattern of its expansion into the archipelago. Reef development within the clastic-free belt probably started earlier in northern Wisconsin than in the Northern Peninsula of Michigan and the Hudson's Bay area. This suggestion of a northeastward spread of the reefs in time is based on Shrock's assignment (1939, p. 534) of the two Manitowoc County reefs, at Grimm and Quarry, to the Cordell member of the Manistique formation. In the Northern Peninsula of Michigan, reefs are unknown in the Manistique deposits and occur first, as far as known, well above the base of the overlying Engadine formation. In the Hudson's Bay area they first appear in the highest Niagaran, in the Attawapiskat beds (Savage and Van Tuyl, 1919).

In the low-clastic belt an expansion southward from northeastern Illinois is better documented. The earliest known reefs in northeastern Illinois occur in the very basal beds of the Waukesha formation near Lemont (Lowenstam, 1948*a*) and in a slightly higher bed within the same formation near Kankakee. The buried structure in the Lemont flagstone quarry, in which the earliest Waukesha reefs occur, suggests that the initial, sporadic reefs may be late Joliet in age (Lowenstam, 1948*a*). Reef growth began in the southwest in the three frontal-reef bastions (McKinley, Bartelso, and Sandoval) and in the more centrally located Marine reef during late St. Clair time, although, in the critical wells, the reef cores are bottomed in the higher Moccasin Springs beds. The top of the St. Clair has been correlated with the Lego-Dixon boundary (Lowenstam, 1949*a*), which in northeastern Illinois lies definitely above the Waukesha-Racine contact and therefore well within the reef-bearing zone. Even if reef-building was initiated in Lego, or even in Waldron, time in the southwesternmost archipelago reefs, the beginning of reef-building here would be much later than that in northeastern Illinois. There is evidence that the reefs spread due southward from Chicago toward the Vermilion County embayment more rapidly than toward the Ozark border. In a well in Sec. 11, T. 18 N., R. 9 E., in Champaign County, the first tongue of rough-water reef outwash occurs 50 feet above a condensed, but typical, Joliet sequence and 250 feet below a 255-foot-thick reef.

A similar pattern of reef migration, from the north-central area toward the southwestern front of the archipelago, is indicated for Indiana. Esarey and Bieberman (1948, pl. 4, fig. 5) have found a subsurface reef in Marion County that started in Osgood time,[4] whereas the outer reef bastions and frontal reefs to the southwest, recently examined by the writer, are Waldron or Lego in age because they show sections underneath the reefs similar to those from Marine, McKinley, and Sandoval in Illinois

[4] Recent examination of the well record by the writer and Mrs. Bieberman showed that this reef section actually bottoms in Laurel deposits.

(Lowenstam, 1949*a*). The front of the reef archipelago in southern Illinois and adjacent southwestern Indiana seems to have been determined by a deep basin, the bottom of which sank below the level of the euphotic ceiling in Moccasin Springs time. The northern limit of the basin was initially a gentle slope, marked by the northern limit of the Moccasin Springs formation across the southern portion of the low-clastic belt. Reefs attaining a foothold near the bottom of the slope gradually incorporated the slope into the shelf by deposition of reef outwash. Movements which later reached a climax in Lower Devonian times accentuated the margin of the basin in which 1,000 feet of sediments ultimately accumulated.

THE PROBLEMS OF LATITUDE AND CLIMATE IN RELATION TO REEF DISTRIBUTION

Intense reef-building in late Niagaran time is known to have extended from the southern parts of Illinois and Indiana to Hudson's Bay (Bell, 1886; Savage and Van Tuyl, 1919) and probably beyond. If the southernmost outlying reef of west-central Tennessee (Wilson, 1949, p. 267) is included, reef-building activity extended from 36° N. to at least 54° N. latitude. Its northernmost limits approach the latitude of the Silurian reefs of Gotland, which lie at about 58° N. The dimensions and density of the reefs in the Hudson's Bay area appear to be similar to those of the southern part of the clastic-free belt of northwestern Illinois and central Wisconsin. Reef-building activity of undiminished vigor occurred at these high latitudes. The published faunal lists of Savage and Van Tuyl (1919) show no impoverishment in genera among either reef-builders or reef dwellers. There is some evidence of climatically controlled faunal zones. The reef-builders, as seen in cores and cuttings from borings on the Marine, McKinley, Bartelso, and Sandoval reefs, consistently show a preponderance of tabulate corals and representatives of *Pycnostylus* over stromatoporoids. This suggests that the stromatoporoids, which are common reef-building elements in the north, diminish southward across the low-clastic belt and are subordinate at its southern border. There is also an increase in the importance of crinoids in the reef debris in the same direction. It is not yet possible to say whether we are dealing with ecologically controlled shifts in biotic composition or with expressions of temperature tolerances, which alone would carry climatic implications. The ecologic aspects are mud tolerance and turbulence. There is evidence that the waters in the frontal archipelago area, near the Ozark island, were both muddier and more turbulent, owing to relatively unobstructed exposure to the prevailing southerly winds. Quantitative paleoecologic studies over the whole reef-bearing area, together with the paleotemperature method of H. C. Urey (1948), a highly promising tool, will enable us to detect and possibly evaluate quantitatively the role of temperature in observed variations in form and biotic composition. Inadequate data on a relatively large territory make it impossible to state with assurance that the temperature anomalies were induced by northward-moving warmer-water currents. Only by such studies as outlined above shall we be able to formulate criteria for coming to grips with the multitude of widely divergent and conflicting speculations that have been advanced to account for the repeated occurrence in the past of apparently subtropical conditions in high latitudes. If we assume that a climatic gradient was established in the

Niagaran and if the isotherms did not coincide with the present parallels of latitude, only then is it permissible to consider migration of the pole and equator, as has been repeatedly suggested (Wegener, 1926) to account for the intense reef-building in the far north and the now established and equally puzzling prevailing southerly winds. A 90° clockwise rotation of the continent would be sufficient to bring the northwestern reefs into a lower latitude suitable for reef growth in a milder climate.

Perhaps there is no need for this radical speculation to account for the northern Niagaran reefs. The southernmost reefs lie only about 4° north of the northernmost reefs of today—32° N. at Bermuda. If a milder world climate is granted—a view supported by the apparent lack of sharp climatic zoning—and if the Niagaran reef-builders had a greater tolerance for lower temperatures (70° F. for reef-building corals today), then we could readily explain the northward spread of the Silurian reefs. The southern known limits of Silurian reefs in the Great Lakes area are determined by environment. South of the known limits of the Niagaran, the Niagaran is either removed by erosion or deeply buried and therefore lacking in outcrops. There is no assurance that reef-building did not take place farther south. Tilting of the earth's axis, to insure the intensity of light required by present-day reef-building organisms, and warm waters are considered by Link (1950, p. 264) in his discussion of Upper Devonian reefs north of the Arctic Circle in northwestern Canada.

A lower threshold of illumination by the Silurian and Devonian reef-builders would permit their extension north of the limits of present-day reef-builders. Although reefs are not found at comparable latitudes today, there are bank-forming, deep- and cold-water corals in the Norwegian fjords (Nordgard, 1929, cited in Yonge, 1940) that justify such a view. If, therefore, bank-building, deep-water corals occur at high latitudes, as at Hudson's Bay, and have a low threshold for light, then reef-building at this latitude does not require such radical changes as tilting of the earth's axis.

The other critical factor is the water temperature. The intense biologic activity required for the mass deposition of calcium carbonate, such as occurs in present-day reefs, cannot take place in relatively cool waters. A lowering of general metabolic activity and of calcium metabolism in particular is inevitable. Because the Niagaran reef-builders erected wave-resistant structures which successfully withstood the impact of wave action, a high rate of metabolic activity is implied. The robust skeletons, the impressive skeletal volume of the reef communities as a whole, and the large contribution of fine bioclastic debris to the expanse of interreef areas further support this view.

The prevailing southerly winds were probably responsible for a constant flow of warm water which moved freely northward through the archipelago. This is of utmost significance.

The extensive evaporite basins of the Salina series are further evidence for a subtropical climate in the area occupied by the reef archipelago. If the Salina is entirely Upper Silurian, as has so far been assumed, then the climate may have become progressively warmer only in post-Niagaran time. It is by no means certain, however, that Salina deposition actually began in post-Niagaran time. Studies of the subsurface Silurian of the Michigan Basin (Cohee, 1948, fig. 9) show that the Niagaran deposits gradually thin radially, so that in southwestern

Michigan and southwestern Ontario they are less than 100 feet thick. Because the Michigan basin has been generally regarded as a negative area of deeper-water deposition during Niagaran times (Cumings and Shrock, 1928), the observed thinning of the sediments was rather unexpected. Cohee believed that the thinning was either due to a local uplift which occurred during the Niagaran, resulting in nondeposition of younger Niagaran deposits, or was the result of uplift and deep erosion between Niagaran and Salina time. Although this interpretation is consistent with the known structural behavior of basins in which sinking may be intermittently interrupted by uplift (both are expressions of the same basic unrest that characterizes active basins), an alternative interpretation must be considered. If the basin remained free of reefs, the adjacent areas might be built up more rapidly by the addition of reef debris, which, if added to the terrigenous clastics, would raise the archipelago bottoms selectively and leave a centrally located depression. This basin, although not solidly walled because of the broken nature of the reef archipelago, would be shielded as a result of its centrally located position within the archipelago. Circulation would have been greatly reduced within the basin through shielding action by the reefs by the time of maximum reef density and maximum buildup of their shallow interconnecting bottoms. In a subtropical climate, hypersaline waters similar to those of the Red Sea could have been produced, initiating evaporite deposition in the basin, while the reefs continued to flourish in the shallow waters of the bordering shelf through Salina time. Thus evaporite deposition may have started during the latter half of the Niagaran, concurrent with the main phase of the archipelago spread, as has also been vaguely suggested by Krumbein, Sloss, and Dapples (1949). Active sinking of the basin would, in time, have further accentuated its depth and would have resulted more readily in the evaporite formation. Then, by the end of Niagaran time, following regional shallowing of the sea, evaporite deposition transgressed the bordering archipelago. Because the Salina beds do not extend beyond the basin-bordering archipelago, except possibly in the southeast, where the Silurian relations to the archipelago border are as yet undefined, the mechanism that may have started salt deposition in the Niagaran certainly must have functioned in the Upper Silurian, as the evaporite deposits are contained within the area built up by reefs and their interreef deposits on the north, west, and south. Biostratigraphic data will be required in the as yet sparingly drilled Michigan basin to determine whether the beginning of evaporite deposition actually took place in Niagaran time. If so, it would constitute another line of evidence of a mild Niagaran climate.

Although the above discussion contributes little to the solution of the problem, the requirements for future research are evident: (1) At present there is no need to postulate such radical hypotheses as continental drift or shift of the earth's axis, inasmuch as certain slight shifts in tolerance of present-day organisms of similar reef-building potential will account for the reef extension to northern areas. (2) Paleoecologic methods can supply the data which will enable us to evaluate the widely divergent speculations which have been made to date.

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REFERENCES CITED

Adams, J. E. (1930) Origin of oil and its reservoir in Yates Pool, Pacos County, Texas: Am. Assoc. Petroleum Geologists Bull. 14, pp. 707–717.

Agassiz, A. (1888) Three cruises of the Blake, vol. 1: Harvard Coll. Mus. Comp. Zoölogy Bull. 14, p. 84.

Arkell, W. T. (1933) The Jurassic system of Great Britain, Oxford, Clarendon Press, 681 pp.

Bather, F. A. (1893) The Crinoidea of Gotland. Pt. 1: The Crinoidea Inadunata: K. svenska vetensk. akad. Handl., vol. 25, no. 2, 200 pp.

Bell, R. (1886) Report on an exploration of portions of the Attawapishkat and Albany rivers: Geol. and Nat. History Survey of Canada Ann. Rept. 2, new ser., pp. 27G–28G.

Bretz, J H. (1939) Geology of the Chicago region: Illinois Geol. Survey Bull. 65, pt. 1, 118 pp.

Chamberlin, T. C. (1877) Geology of eastern Wisconsin; in: Geology of Wisconsin, vol. 2, Madison, Wisconsin Geol. Survey, pp. 360–371.

Cohee, G. (1948) Thickness and lithology of Upper Ordovician and Lower and Middle Silurian rocks in the Michigan basin: U.S. Geol. Survey Oil and Gas Inv. Preliminary Chart 33, sheet 2.

Cooper, G. A., *et al.* (1942) Correlation of the Devonian sedimentary formations of North America: Geol. Soc. America Bull. 53, pp. 1729–1788.

Cumings, E. R. (1932) Reefs or bioherms? Geol. Soc. America Bull. 43, pp. 331–352.

——— and Shrock, R. R. (1928) The geology of the Silurian rocks of northern Indiana: U.S. Dept. Cons. Pub. 75, 226 pp.

Emery, K. O.; Tracey, T. I.; and Ladd, H. S. (1949) Submarine geology and topography in the northern Marshalls: Am. Geophys. Union Trans., vol. 30, pp. 55–58.

Esarey, R. E., and Bieberman, D. F. (1948) Correlation of the Waldron and Mississinewa formations: Indiana Dept. Cons., Div. Geology Bull. 3, 36 pp.

Fenton, C. L. (1931) Niagaran stromatoporoid reefs in the Chicago region: Am. Midland Naturalist, vol. 12, pp. 203–212.

Foerste, A. F. (1935) Correlation of Silurian formations in southwestern Ohio, southeastern Indiana, Kentucky, and western Tennessee: Denison Univ., Sci. Lab., Bull. 30, pp. 119–204.

Freeman, L. B. (1949) Regional aspects of Cambrian and Ordovician subsurface stratigraphy in Kentucky: Am. Assoc. Petroleum Geologists Bull. 33, pp. 1655–1681.

Grabau, A. W. (1903) Paleozoic coral reefs: Geol. Soc. America Bull. 14, pp. 337–352.

——— (1924) Principles of stratigraphy, New York, A. G. Seiler, 1185 pp.

Hadding, A. (1941) The pre-Quaternary sedimentary rocks of Sweden. VI. Reef limestones: Lunds Geol.-min. Inst. Meddelanden no. 90, 137 pp.

Hall, James (1862) Physical geography and general geology: Wisconsin Geol. Survey Rept., vol. 1, pp. 1–72.

Krempf, A. (1930) About the shape of the coral-reefs and the system of the alternating winds: Fourth Pacific Sci. Cong., vol. 2A, pp. 477–480.

Krumbein, W. C.; Sloss, L. L.; and Dapples, E. C. (1949) Sedimentary tectonics and sedimentary environment: Am. Assoc. Petroleum Geologists Bull. 33, pp. 1859–1891.

Ladd, H. S. (1944) Reefs and other bioherms: Rept. Comm. Marine Ecology as related to Paleontology, 1943–44, National Research Council, no. 4, pp. 26–29.

——— (1950) Recent reefs: Am. Assoc. Petroleum Geologists Bull. 34, pp. 203–214.

Lecompte, M. (1938) Quelques types de récifs Siluriens et Dévoniens de l'Amérique du Nord. Essai de comparaison avec les récifs coralliens actuels: Mus. royale histoire nat. Belgique Bull. 14, 51 pp.

Line, T. A. (1950) Theory of transgressive and regressive reef (bioherm) development and origin of oil: Am. Assoc. Petroleum Geologists Bull. 34, pp. 263–294.

Lowenstam, H. A. (1948*a*) Biostratigraphic studies of the Niagaran inter-reef formations in northeastern Illinois: Illinois State Mus. Sci. Papers, vol. 4, 146 pp.

——— (1948*b*) Marine Pool, Madison County, Illinois, Silurian reef producer; in: Structure of typical American oil fields, vol. 3, Am. Assoc. Petroleum Geologists, pp. 153–188.

——— (1949*a*) Niagaran reefs in Illinois and their

relation to oil accumulation: Illinois Geol. Survey Rept. Inv. 145, 36 pp.

——— (1949b) Facies analyses of the Niagaran rocks in Illinois: Illinois Acad. Sci. Trans. 42, pp. 113–115.

——— and DUBOIS, E. P. (1946) Marine Pool, Madison County, a new type of oil reservoir in Illinois: Illinois Geol. Survey Rept. Inv. 114, 30 pp.

MA, T. Y. (1937) On the seasonal growth in Palaeozoic tetracorals and the climate during the Devonian period: Palaeontologia Sinica, ser. B, vol. 2, no. 3, pp. 1–96.

OTTER, G. W. (1937) Rock-destroying organisms in relation to coral reefs: Great Barrier Reef Expedition, 1928–1929, British Mus. Nat. History Sci. Repts., vol. 1, pp. 353–391.

ROLL, A. (1934) Form, Bau und Entstehung der Schwammstotzen in süddeutschen Malm: Zeitschr. Palaeontologie, vol. 16, pp. 197–246.

SAVAGE, T. E. (1926) Silurian rocks in Illinois: Geol. Soc. America Bull. 37, pp. 513–533.

——— and VAN TUYL, F. M. (1919) Geology and stratigraphy of the area of Paleozoic rocks in the vicinity of Hudson and James bays: Geol. Soc. America Bull. 30, pp. 339–377.

SCHMIDT, H. (1935) Die bionomische Einteilung der fossilen Meeresböden: Fortschr. Geologie u. Palaeontologie, vol. 12, no. 38, 154 pp.

SHROCK, R. R. (1939) Wisconsin Silurian bioherms (organic reefs): Geol. Soc. America Bull. 50, pp. 529–562.

STOCKDALE, P. B. (1931) Bioherms in the Borden group of Indiana: Geol. Soc. America Bull. 42, pp. 707–718.

SVERDRUP, H. V.; JOHNSON, M. W.; and FLEMING, R. H. (1942) The oceans, their physics, chemistry, and general biology, New York, Prentice-Hall, Inc., 1087 pp.

UMBGROVE, J. H. F. (1930) The influence of the monsoons on the geomorphology of coral islands: Fourth Pacific Sci. Cong. Proc., vol. 2A, pp. 49–54.

——— (1949) Structural history of the East Indies, Cambridge, Cambridge University Press, 63 pp.

UREY, H. C. (1948) Oxygen isotopes in nature and in the laboratory: Science, vol. 108, pp. 489–496.

VAUGHAN, T. W. (1911) Physical conditions under which Paleozoic coral reefs were formed: Geol. Soc. America Bull. 22, pp. 238–252.

——— (1919) Corals and the formation of coral reefs: Smithsonian Inst. Ann. Rept. for 1917, pp. 189–276.

WALTHER, T. (1893–1894) Einleitung in die Geologie als historische Wissenschaft. 3. Lithogenesis der Gegenwart, Jena, Verlag Gustav Fischer, pp. 535–1055.

——— (1927) Allgemeine Palaeontologie, Berlin, Verlag Gebrüder Borntraeger, 809 pp.

WEGENER, A. (1926) Palaeographische Darstellung der Theorie der kontinental Verschiebungen; in: DACQUÉ, E., Paläogeographie: Enzyklopädie der Erdkunde, Leipzig und Wien, Franz Deuticke, 196 pp.

WILLIS, B. (1912) Paleogeographic maps—Ordovician and Silurian; in: Outlines of geologic history with especial reference to North America, Chicago, University of Chicago Press, 306 pp.

WILSON, C. W., JR. (1949) Pre-Chattanooga stratigraphy in central Tennessee: Tennessee Dept. Cons., Div. Geology, Bull. 56, 407 pp.

YONGE, C. M. (1940) The biology of reef-building corals: Great Barrier Reef Expedition, 1928–1929, British Mus. Nat. History Sci. Repts., vol. 1, pp. 353–391.

COMMUNICATIONS AND ANNOUNCEMENTS

WYOMING GEOLOGICAL ASSOCIATION FIFTH ANNUAL FIELD CONFERENCE

The fifth annual field conference of the Wyoming Geological Association will be held in southwestern Wyoming on August 9, 10, 11, as announced by John F. Partridge, district geologist for the California Company and general chairman of the field trip committee. The conference will be headquartered in Kemmerer, Wyoming, for the entire time; but daily field trips, ranging from the Uinta Mountains on the south to the Wind River on the north, will allow the group to study the geology of a wide area of the Green River Basin and western Wyoming overthrust belt.

John Harrison, of the Argo Oil Corporation, is editor of the 1950 *Guidebook*, and D. A. Moore, of the Superior Oil Company, has been named trip chairman in charge of actual field arrangements. Participants will register in Kemmerer the afternoon and evening of August 8.

R. W. Mallory, of the Stanolind Oil and Gas Company, Box 40, Casper, is secretary of the Wyoming Geological Association.

ANNOUNCEMENT OF NEW JOURNAL

A new geochemical journal, *Geochimica et cosmochimica acta*, is being published by Butterworth-Springer, 436 Strand, London, W.C. 2, England, under the editorship of Professor L. R. Wager. The journal is international in scope, and the plan is to give prompt publication to geochemical papers from all over the world. They will be published in English, French, or German, with English summaries. Authors are to be given 75 reprints free of charge. Manuscripts should be sent to Earl Ingerson, U.S. Geological Survey, Washington 25, D.C., who is American editor for the new publication.

Subscription prices have not yet been announced, but the American agent is the Academic Press, Inc., 125 East Twenty-third Street, New York 10, New York, to whom inquiries should be addressed.

INTERNATIONAL UNION OF CRYSTALLOGRAPHY: SECOND INTERNATIONAL CONGRESS AND GENERAL ASSEMBLY, 1951

The Executive Committee of the International Union of Crystallography has accepted an invitation from the Swedish national committee to hold the Second International Congress on Crystallography and the Second General Assembly of the Union in Stockholm from June 27 to July 3, 1951. These dates have been chosen in consultation with the Swedish national committee and with the national committees of all the countries which adhere to the Union. It is hoped that this early notice will make it possible for those interested in crystallography and its related fields to arrange to attend. Further information can be obtained from the general secretary of the Union, Dr. R. C. Evans, Cavendish Laboratory, Cambridge, England.

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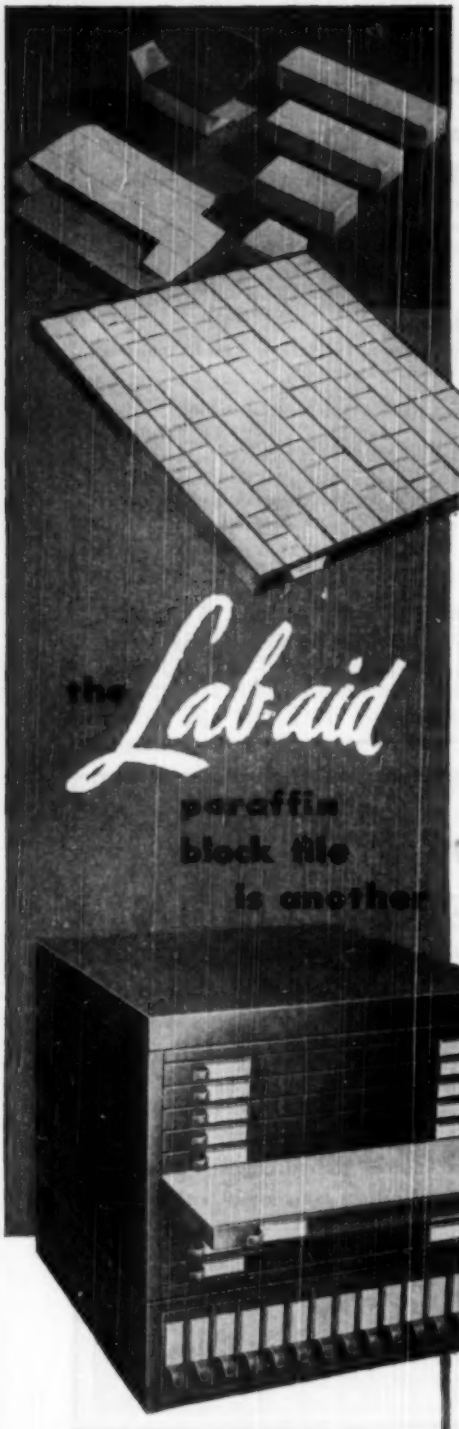

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